

SEDIMENTOLOGY OF AN UPPER CAMBRIAN FLYSCH-PARALIC SEQUENCE (DENISON GROUP)
ON THE DENISON RANGE, SOUTHWEST TASMANIA

Being a detailed study of a proximal flysch formation (Singing Creek Siltstone) and of part of a fluviomarine sandstone formation (Great Dome Sandstone) combined with a regional study of Lower Palaeozoic stratigraphy and tectonics in the Adamsfield-Florentine area.

by

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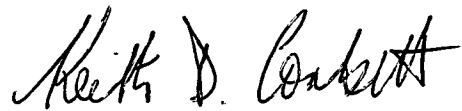
Thesis submitted in partial fulfilment
of the requirements for the degree of
Doctor of Philosophy.

UNIVERSITY OF TASMANIA

HOBART

July, 1970.

This thesis contains no material which has been accepted for the award of any other degree or diploma in any university, and to the best of my knowledge and belief contains no copy or paraphrase of material previously published or written by another person, except when due reference is made in the text.

A handwritten signature in black ink, reading "Keith D. Corbett". The signature is written in a cursive style with a large, stylized "K" and "C".

Keith D. Corbett.

CONTENTS

	page number
LIST OF FIGURES	
LIST OF PLATES	
SUMMARY	
INTRODUCTION	
Nature of Study	1
Location and Access	2
Geomorphology	4
Previous Literature	5
Acknowledgements	6
Specimens and Thin Sections	7
CHAPTER 1 - GEOLOGICAL SETTING - A BRIEF ACCOUNT OF THE TECTONICS AND STRATIGRAPHY OF THE ADAMSFIELD AREA	
A. MAJOR PALAEOZOIC ELEMENTS OF WESTERN TASMANIA	8
1. Precambrian Elements	8
2. The Cambrian Elements	11
3. Upper Cambrian-Lower Ordovician Elements -- the Owen-type Conglomerates	13
4. The Ordovician to Devonian Elements	14
5. Tabberabberan Elements	14
B. TECTONICS OF THE ADAMSFIELD-FLORENTINE AREA	
1. Structures of the Tyennan Geanticline	15
2. Structures of the Jubilee Block	15
3. Tectonics of the Adamsfield Trough	
a. General	17
b. Boyd Basin	20
c. Stepped Hills-Clear Hill Basin	21
d. Denison Range Basin	21
e. Battlement Hills Basin	23
f. Cambrian Orogenic Movements	23
4. Tabberabberan Structures	28
C. STRATIGRAPHY	
1. Older Precambrian of Tyennan Geanticline	29
2. Younger Precambrian of Jubilee Block	29
3. Cambrian Unfossiliferous Sequences and Igneous Rocks	
a. Older Cambrian of Boyd Basin	31
(i) Wedge River Beds	31
(ii) Conglomerates overlying Wedge River Beds	32
(iii) Turbidite sequence near Mt. Wedge	33
(iv) Greywacke-chert-mudstone sequence	34
(v) Igneous rocks of Boyd Basin	35

	page number
b. Mt. Mueller Area	36
c. Clear Hill Area	37
d. Trial Ridge Beds	37
e. Major Ultramafic Bodies	38
4. Cambrian Fossiliferous Sequences	40
a. Scotts Peak Road Sequence	41
b. Adamsfield Beds	42
c. Denison Group	44
(i) Singing Creek Siltstone	46
(ii) Great Dome Sandstone	47
(iii) Denison Group in other areas	48
5. Ordovician Stratigraphy	
a. General	49
b. Reeds Conglomerate	50
c. Tim Shea Sandstone	53

CHAPTER 2 - SEDIMENTOLOGY OF THE SINGING CREEK SILTSTONE

A. GENERAL	55
B. ARRANGEMENT OF ROCK TYPES	55
C. GROUP A - SILTSTONES	58
1. Lamination and Graded Bedding	58
2. Deformational Structures	60
3. Sole Marks	60
4. Petrology	60
5. Deposition of the Siltstones	62
D. GROUP B - THIN-BEDDED SANDSTONES	
1. General Features	64
2. Stratification	65
a. Bed Thickness	65
b. Types of Stratification Surface	66
3. Examples of Lateral Variations	68
4. Sole Marks	72
a. Current Marks	73
(i) Longitudinal ridges and furrows	73
(ii) Obstacle scours	74
(iii) Tool marks	74
(iv) Flute marks	74
(v) Channels	76
b. Load Structures	76
5. Graded Bedding	78
6. Internal Structures	79
a. Basal Structureless Division	79
b. Division of Parallel Lamination	80
c. Cross-Lamination and Ripple Mark	82
d. Deformed Lamination	83
e. Upper Division of Laminated Siltstone	85

	page number
7. Composition and Provenance	85
a. Modal Analyses	86
b. Compositional Components	87
c. Provenance	89
8. Texture	89
a. Matrix	89
b. Grainsize Analysis and Sorting of Sand Fraction	91
(i) Method	91
(ii) Sorting	93
(iii) Significance of sorting	94
c. Rounding	96
9. Classification of the Sandstones	96
10. Mode of Emplacement of the Sandstones	
a. Background	99
b. Some Evidence from Sedimentary Features	102
(i) Lack of sorting and lamination in basal division	103
(ii) Topographic control of deposition	103
(iii) Transport without deposition (auto-suspension)	103
(iv) Relationship to Group C sandstones and Group D conglomerates	104
 E. GROUP C -- THICK-BEDDED SANDSTONES	
1. General Features	105
2. Sole Marks	106
3. Internal Structures	107
a. Contorted Lamination	107
b. Wavy Lamination or Dish Structure	108
c. Parallel Lamination	108
d. Ripple Mark	109
e. Large-scale Cross-bedding	109
f. Elutriation Columns	110
4. Graded Bedding	110
5. Lateral Variations	111
6. Composition and Texture	111
7. Previous Descriptions of Similar Sandstones	112
8. Origin of Dish Structure	113
 F. GROUP D - CONGLOMERATES	114
1. Distribution	114
2. Composition	115
3. Texture	115
4. Coarse Conglomerates North of Flagstone Knoll	117
5. Ridge Above Boyes River	117
6. Head of Singing Creek	118
7. Mode of Emplacement of the Conglomerates	123
a. Background	123
b. Sedimentary Features	125
(i) Association with Group B sandstones	125
(ii) Poor sorting	125
(iii) Upper sandy division with flat lamination	125

	page number
(iv) Distribution of sand and gravel within beds	126
(v) Deformational structures	126
(vi) Topographic control of deposition	126
c. The Coarse Conglomerates	127
d. Transport by Antidune Flows	127
G. RECONSTRUCTED NATURE OF THE DEPOSITING CURRENTS	130
H. GROUP E - SLUMP SHEETS	132
1. General	132
2. Stages of Deformation	133
a. Coherent Stage	135
b. Semi-coherent Stage	135
c. Incoherent Stage	136
d. Homogeneous Stage	137
3. Folding	
a. General	138
b. Isoclinal Folds	139
c. Irregular Folds	142
d. Box-type Folds	142
e. Broad Synclines and Complex Anticlines	142
f. Parasitic Folds and Slump Lineation	143
4. Basal Contacts	143
5. Upper Contacts	145
6. Directional Properties	146
a. Azimuth of Movement - Fold Axis Orientation	146
b. Sense of Movement	148
7. Genesis of the Slump Sheets	149
8. Palaeoslope-Palaeocurrent Relationship	149
I. DEPOSITIONAL ENVIRONMENT OF THE SINGING CREEK FORMATION	
1. Major Features of the Deposit	152
2. Ancient and Modern Analogues	152
3. Conclusions	157
CHAPTER 3 - SEDIMENTOLOGY OF THE GREAT DOME SANDSTONE	
A. INTRODUCTION	159
B. BASAL TRANSITION BEDS	161
C. CYCLIC FACIES	162
1. Definition of Cycles	162
2. Lithofacies A. Thick-Bedded Coarse Sandstone Units	166
3. Lithofacies B. Thin-Bedded Sandstone and Siltstone with Abundant Worm Burrows	169
4. Lithofacies C. Siltstone	171
5. Lithofacies D. Sandstone with Low-Angle Cross-Bedding	172
6. Lithofacies E. Cross-Bedded Sandstone with Marine Fossils	176
7. Origin of the Cycles	181

	page number
D. UPPER PART OF THE FORMATION	181
CHAPTER 4 - SUMMARY OF DEPOSITIONAL HISTORY OF DENISON RANGE BASIN	183
BIBLIOGRAPHY	189-207
APPENDIX A -- ORDOVICIAN STRATIGRAPHY	A1-A18
APPENDIX B - CATALOGUE OF ROCK SPECIMENS AND THIN SECTIONS	B1-B10

LIST OF FIGURES

	page number
Figure 1 Palaeozoic Elements of Western Tasmania	9
Figure 2 Geology of the Adamsfield-Florentine Area (Interpretive)	16
Figure 3 Structural Elements of Adamsfield Trough and Adjacent Areas	18
Figure 4 Diagrammatic Section Showing Cambrian Unconformi- ties in Adamsfield Trough - Tim Shea Area	24
Figure 5 Geology of Central and Northern Denison Range	45
Figure 6 Block Diagram Showing Distribution of Reeds Conglomerate and Tim Shea Sandstone	51
Figure 7 Singing Creek Formation. Composite Section Flagstone Knoll Area	56
Figure 8 Scale Section Showing Variations in Three Sandstone Beds	70
Figure 9 Size Distribution of Quartz Grains in Sandstones and Conglomerates	92
Figure 10 Scale Section of Three Conglomerate Beds	120
Figure 11 Detailed Section Through Part of Singing Creek Formation	134
Figure 12 Suggested Development of Isoclinal Fold Pattern in Slump No.14	140
Figure 13 Section Through Cyclic Facies of Great Dome Sandstone	160
Figure 14 Detailed Columnar Section Through Part of Cyclic Facies	164
Figure 15 Description and Interpretation of Complete Cycle from Cyclic Facies of Great Dome Sandstone	165
Figure 16 Evolution of the Denison Range Basin	184
Figure 17 Palaeocurrents - Denison Range Basin	186

TABLE

Table 1 follows page 86

LIST OF PLATES

		follows page number
PLATE 1	A. Reeds Conglomerate on western slope of Stepped Hills B. Upper Sandstone Member of Reeds Conglomerate at Battlement Hills	51
PLATE 2	A. Typical outcrop of Reeds Conglomerate at Clear Hill B. Reeds Conglomerate, northern Denison Range	52
PLATE 3	A. Red facies of Tim Shea Sandstone, near crest of Tim Shea B. Grey facies of Tim Shea Sandstone, north of Mt. Mueller	53
PLATE 4	A. Graded sandy layers within Group A siltstones B. Cross-laminated and ripple-marked layers within Group A siltstones	58
PLATE 5	A. Graded quartzose laminae in Group A siltstones B. Photomicrograph of contact between sandy and silty laminae	60
PLATE 6	A. Cross-laminated and ripple-marked laminae within Group A siltstones B. Composite sandstone sequence typical of Group B, with numerous stratification surfaces	61
PLATE 7	A. Group B sandstones B. Compositely-bedded sandstone sequence showing profusion of stratification surfaces	65
PLATE 8	A. Sandstone layer between two slump sheets B. Same unit as above, about 100 feet further south	66
PLATE 9	A. Sandstone layer between two incoherent slump sheets B. Sandstone layer with truncated flute marks	67
PLATE 10	A. Sandstone layer with truncated flute marks at base B. Load pockets filled with coarse grit developed at contact between two sandstone beds	68
PLATE 11	A. Sandstone layer showing thickness variation due to filling of large channel structure B. South end of outcrop shown in Fig. 8	69
PLATE 12	A. Conglomerate lens of Bed A passes laterally into thin sand lamina B. South end of outcrop shown in Fig. 8	70
PLATE 13	A. Load-deformed flute marks on base of sandstone layer separating two incoherent slump sheets B. Flute marks filled with cross-laminated fine sand	73
PLATE 14	A. Sole of bed showing longitudinal ridge and furrow structure modified by pre-existing transverse ripples B. Small isolated flute marks on sole of sandstone bed	74

PLATE 15	A. Sandstone filling small channel structure cut in incoherent slump sheet B. Sole of sandstone bed showing superimposed load marks with preferred orientation	75
PLATE 16	A. Large load structures on sandstone bed overlying small slump sheet B. Series of load structures along sandstone overlying small slump sheet	76
PLATE 17	A. Outcrop showing folded lamination in upper silty part of graded sandstone bed B. Detail of folded lamination	83
PLATE 18	A. Photomicrograph of typical Group B sandstone B. Photomicrograph showing introduction of dark sericitic matrix from flame structure into base of sandstone bed C. Conglomeratic part of sandstone D. Typical fine conglomerate of Group D	87
PLATE 19	A. Composite sequence of Group C sandstones on crest of Flagstone Knoll B. Detail of above, showing one complete bed with dish structure and parts of two others	106
PLATE 20	A. Single Group C sandstone bed showing contorted structure at base merging upwards into dish structure B. Same bed as above, 10 feet to south	107
PLATE 21	A. Composite sequence of mainly Group C sandstone beds B. Photomicrograph of elutriation column in Group C sandstone	109
PLATE 22	A. Graded conglomerate beds two miles north of Flagstone Knoll B. Upper surface of graded conglomerate bed showing projecting quartzite boulder near hammer	116
PLATE 23	A. Graded fine conglomerate bed, ridge above Boyes River B. Poorly-graded fine conglomerate bed with "floating" pebbles	117
PLATE 24	A. Composite sequence of conglomerate and sandstone, ridge above Boyes River B. Conglomerate-sandstone zone near head of Singing Creek	118
PLATE 25	Upper part of zone shown in Plate 24B. Conglomerate bed with large deformational sole structures affecting underlying sandstones	119
PLATE 26	A. Beds 2 and 3 of Fig. 10 B. Approximately same horizon as above, about 100 feet to north	121

		follows page number
PLATE 27	. Photo series showing 20-foot section of Slump No.14	135
PLATE 28	A. Complex coherent deformation in part of Slump No.14 B. Semi-coherent deformation in Slump No.17	136
PLATE 29	A. Semi-coherent deformation in Slump No.28, with large synclines and complex anticlines B. Typical incoherent slump sheet with irregular sandstone remnants and "floating" pebbles	137
PLATE 30	A. Incoherent slump sheet (No.20) with flow cleavage, overlain by sandstone raft showing deformed flute mark B. Homogeneous slump sheet (No.2) with horizontal flow cleavage	138
PLATE 31	A. Large conjugate box-type fold in semi-coherent slump sheet (No.22) B. Sandstone rafts folded into top of slump sheet (No.16) and truncated by overlying sandstone bed	142
PLATE 32	A. Outcrop of Transition Beds at base of Great Dome Sandstone, Staircase Rocks B. Large composite flow-roll, Transition Beds	161
PLATE 33	A. View of Cyclic Facies of Great Dome Sandstone in type area at Staircase Rocks B. Thick-bedded unit of Lithofacies A disconformably overlying thin-bedded zone	162
PLATE 34	A. Lithofacies B. Flow-rolls of laminated sandstone surrounded by dark sandy siltstone B. Bioturbated sandstone of upper part of Lithofacies B (inter-tidal sand flats ?)	169
PLATE 35	A. Dark siltstone of Lithofacies C, with intercalations of laminated fine sandstone B. Rafts and flow-rolls of laminated Lithofacies D sandstone surrounded by sandy siltstone of Lithofacies C	171
PLATE 36	A. Erosional base of main upper sandstone unit of Lithofacies D in Cycle 6 B. Low-angle cross-bedding in Lithofacies D sandstone of Cycle 2	173
PLATE 37	A. Typical cross-bedded sandstone of Lithofacies E, Cycle 1 B. Moulds of current crescents around worm tubes on underside of sandstone bed, Lithofacies E	176
PLATE 38	A. Typical cross-bedded sandstones and fine conglomerates of upper part of Great Dome Sandstone near Reeds Peak B. North end of Denison Range showing transition from the Cyclic Facies below to red sandstones of upper part of Great Dome Sandstone (centre) to the thick-bedded conglomerates of Reeds Conglomerate on crest.	181

SUMMARY

The Denison Range Basin is one of several small basins along the Adamsfield Trough, a Cambrian eugeosynclinal structure developed between a geanticline of "Older" Precambrian metasediments to the west (Tyennan Geanticline) and a complex belt of mainly "Younger" Precambrian rocks to the southeast (Jubilee Block). The trough is overlapped from the east by marine Ordovician rocks of the Florentine Synclinorium. Most of the filling of the Denison Range Basin took place in the Upper Cambrian, after the main igneous activity in the trough (including the emplacement of ultramafics onto the seafloor) had ceased, and the sediments record a transition from flysch-type turbidites and siltstones to molasse-type sandstones and fanglomerates.

The basin-filling consists of four units, viz. the Trial Ridge Beds (2000 feet) of mainly conglomerates and turbidites at the base, followed with low-angle unconformity by the Singing Creek Siltstone (2600 feet) of turbidites and siltstones, followed conformably by the Great Dome Sandstone (1700 feet) of mainly shallow marine and fluvial sandstones, followed conformably by the non-marine Reeds Conglomerate (5200 feet). All the rocks were apparently derived from the Precambrian quartzites and schists of the Tyennan Geanticline, such that the sandstones are very quartzose (including the flysch-types, which are pure quartz wackes) and the siltstones are very micaceous. The two middle formations constitute the Denison Group, and have been studied in detail.

The Singing Creek Siltstone consists of four main rock types, viz. laminated siltstone - fine sandstone (Group A); thin-bedded sandstones which are usually graded (Group B); thick-bedded sandstones which are commonly non-graded and which may show dish-structure (Group C); and quartzose pebble- to boulder-grade conglomerate (Group D). Most of the conglomerates and thick sandstones occur in composite sequences up to 100 feet or so thick, forming what appear to be large lenses or channels surrounded by siltstones. The field evidence strongly suggests that all the rock types were deposited by the same kind of current, the variations within the current causing different grainsizes to be deposited in different places. Lateral variations from conglomerate

to sandstone to siltstone occur within single beds, and there are abundant intermediate types between the four groups.

The depositing currents were: (i) spasmodic and waning, as indicated by the graded bedding and sequence of internal structures; (ii) bottom-hugging, as indicated by the influence of small bottom features on deposition; (iii) capable of carrying pebbles and in some cases even boulders in suspension; (iv) probably capable of auto-suspension, since the presence of isolated depressions filled with coarse grains beneath many beds indicates that the gravel-bearing phase of the current had passed without depositing (except in the small hollows). The currents can be reconstructed as sediment-laden density underflows (i.e. turbidity currents) in which the bulk of the coarse material was carried in a dense frontal and basal part where flow conditions were in the high upper flow regime (antidunes and above). Suspension of the coarse material may have been largely by dispersive pressure rather than turbulence, considering the large grainsizes and probable high concentrations involved. Deposition from this part was spatially restricted, probably because the flows were channelized, and resulted in thick, laterally-variable layers of sand and gravel (Group C sandstones and Group D conglomerates). Antidune lamination (dish structure) was formed when the load consisted mainly of fine sand, but otherwise the bulk of the deposit tended to be structureless and unsorted (Bouma's division a).

Deposition was slower as the current velocity declined to the plane bed phase of the upper flow regime, and a thin traction carpet was formed in which the sediment was sorted to some extent and flat laminae were formed (division b). Some reworking and smoothing-off of the previously-deposited sediment probably occurred at this stage. The plane bed was replaced by ripples (division c) as the current declined rapidly through the lower flow regime. Deposition from these phases was more widespread than from the early part and less influenced by bottom topography, producing thin laminated sandstone layers (many Group B beds) as well as laminated tops to the earlier deposits. In many cases, laminated sand was laid down directly on the flute-marked surface which had been scoured but left uncovered by the frontal part of the current.

The dense initial part of the current was apparently flanked and followed by a more diffuse part carrying mainly silt and fine sand, probably in turbulent suspension. Deposition from this phase was blanket-like and widespread (Group A), probably because the higher part of the flow was able to overtop the channel and spread over surrounding areas. Much of the silt deposited on the initial sand and gravel deposits was apparently eroded by later flows following the same path, so that the main areas of silt accumulation were marginal to the coarse deposits.

Open-cast (truncated) slump sheets are very abundant in parts of the section, and range from simply-folded coherent types to completely disaggregated and mixed types with a prominent flow cleavage. Fold axes in the coherent slumps at Flagstone Knoll are strongly aligned and indicate movement at right angles to the palaeocurrent direction. The slump sheets are interpreted as local features initiated by contemporaneous movements on cross-faults, and several such faults are present in the area. The formation is interpreted as a submarine fan complex, and the pattern of coarse-grained lenses in a matrix of siltstones resembles the channel-interfluvial arrangement of modern and Tertiary fans.

The overlying Great Dome Sandstone shows a cyclic arrangement of lithofacies in its lower half. A detailed section through 200 feet of the Cyclic Facies shows six cycles, most of which consist of five lithofacies as follows. At the base, one or several cross-bedded coarse sandstone units with an erosional base, interpreted as distributary channel deposits. These are interbedded with and overlain by thin-bedded sandstones and siltstones showing abundant bioturbation, interpreted as mainly tidal-flat deposits. Above these is a siltstone unit with interbeds of fine sandstone, interpreted as lagoonal muds interfingering with barrier-back sands. The sandstones show low-angle cross-bedding and become more predominant upwards, the upper unit possibly being a barrier beach. The upper lithofacies of most cycles consists of irregularly cross-bedded sandstones containing marine gastropods, interpreted as surf-zone deposits. The cycles are best explained in terms of the delta-switching process, whereby the distributary system migrates to a different part of the delta complex, allowing a temporary marine

incursion as the abandoned area subsides. The upper part of the formation consists mainly of channel deposits. Final filling of the basin occurred under non-marine conditions, with the deposition of gravel and sand by braided streams on large alluvial fans to form the Reeds Conglomerate.

INTRODUCTION

Nature of Study

This study presents a detailed analysis of an Upper Cambrian flysch-paralic sequence (Denison Group) on the Denison Range, combined with a regional study of the Cambrian trough in which this sequence was deposited. Most of the work concerns the well-exposed flysch sequence (Singing Creek Siltstone), and it is believed that several important insights into the nature of turbidity currents on submarine fans have been gained. The overlying sequence of cyclic paralic sediments (Great Dome Sandstone) is also described, although in somewhat less detail because of space and time considerations. The cycles have been interpreted in terms of alternations of delta distributary-tidal flat sedimentation with shallow marine incursions.

The study was initiated when the author was granted a Broken Hill Proprietary Postgraduate Scholarship in late 1965, for which field work commenced in 1966. The conditions of the scholarship required that the recipient should work on, and produce a map of, part of the mineral exploration lease held by the B.H.P. over southwest Tasmania at that time. The Adamsfield area was chosen because of the author's interest in the Lower Palaeozoic, and because while access was not good it was at least possible to drive to the area at any time and to walk to most points within one to two days. No specific problem within the field of the Lower Palaeozoic succession was outlined for study at the beginning, since the geology of the area was poorly known. For this reason it was decided to map the area from north to south, while helicopter transport was available, and to delay detailed work until a suitable problem emerged.

The discovery of the remarkable Upper Cambrian sequence on the Denison Range solved the problem of what to study in detail, but the mapping program was continued in order to determine the regional setting of this sequence and to satisfy the scholarship requirements. Much more work remains to be done, however, particularly in the southern part of the area where the mapping is essentially of a reconnaissance nature only.

Apart from the regional study, the main part of the work involved detailed sedimentology, particularly the origin and significance of sedimentary structures. This is a relatively new field of geology, at least insofar as most of the major advances have occurred in the last 10-15 years, and was not taught as a special subject at this department. The author was forced, therefore, to learn the subject from the beginning, and most of the study has been done without any specialized supervision. A great deal more detailed work could be done on the Denison Range sequence, since it appears to be one of the best exposed and most informative sequences of its kind in the world.

Location and Access

The old osmiridium-mining township of Adamsfield is located about 50 miles WNW of Hobart, and lies in the northeastern part of Tasmania's rugged southwest district. The township sprang up after an osmiridium "rush" in 1925, but has been virtually a ghost town since 1940. The nearest permanent settlement is the logging township of Maydena (Fig. 1), about 15 miles to the ESE, which can be reached via a sealed highway from Hobart. The recently-constructed Hydro-Electric Commission village of Strathgordon lies about 17 miles to the west on the Gordon Road. The area mapped is uninhabited at present.

The building of the Gordon Road by the H.E.C. from Maydena to Strathgordon, in the period 1962-1967, has greatly aided access to the area, but Adamsfield itself is still only accessible with four-wheel drive vehicles by tracks along the Saw Back Range and across the Florentine Valley from Tim Shea (Fig. 2). Access to the areas north of this is by foot or helicopter only. The Scotts Peak Road branches southward from the Gordon Road near the head of the Florentine River (Fig. 2), and provides vehicular access to the Scotts Peak area for the first time. Most of the mapping of the northern areas was facilitated by helicopter transport provided by the B.H.P. Company, using the H.E.C. helicopter base at Hermit Camp on the Gordon Road. The present study would not have been possible without this assistance.

The area mapped covers the major part of the Tasmanian Lands and Surveys Wedge Sheet (one inch to one mile) plus the southeastern corner of the Nive Sheet. Both these are contoured at 50-foot intervals, and a prepared composite was used as a base map (Appendix C). The smaller component sheets, at a scale of 40 chains to one inch, were used for detailed mapping of some areas. An excellent coverage of aerial photographs at various scales was available, and data plotted on these could be transferred directly to the contoured base maps. Much of the bedrock in the area is covered by superficial deposits, consisting mainly of Quaternary alluvium, talus, and glacial and glacio-fluvial deposits. These are shown fairly fully on the Appendix C map, while Fig. 2 of the thesis is an interpretive map of the bedrock geology. Previous work by the author in the Florentine Valley (Corbett, 1964) is incorporated in these maps.

Field work in southwest Tasmania poses special problems because of the rugged topography and harsh climate. The average rainfall ranges from 100-140 inches per year, and the cold wet winters are not suitable for outdoor work except near permanent camps. The short summers (December-March), although generally mild, are subject to rapid changes of weather and sudden storms, and care must be exercised when working from tent camps. Most of the author's work was carried out from a tent camp on the Denison Range, using one field assistant. Periods of 3-4 weeks were spent in the field, alternating with short rest periods in Hobart to replenish supplies. A total of nearly 12 months was spent in the field from January 1966 - June 1969.

Weather problems are accentuated at the high altitudes, and field work was possible on less than half the days actually spent in the field. The fact that the range was accessible only for a short period each year made it impossible to check any theories and ideas arising from literature study and theoretical considerations until the following summer, and this has considerably hampered the sedimentological analysis. Because of this, considerable use had to be made of photographs of outcrops and structures.

Geomorphology

The area lies within the "fold structure province" of western Tasmania, in which physiographic features are largely controlled by Lower Palaeozoic fold trends (Davies, 1965). The "fault-structure province" of central Tasmania, developed on the sub-horizontal Permo-Trias sediments and Jurassic dolerite, forms the eastern limit of the area, with remnants on Mt. Mueller and Mt. Wedge (Fig. 2). Four major physiographic units are present from west to east, viz. the Precambrian ridges of the Tyennan Geanticline, the irregular area of Cambrian rocks in the Boyd Valley, the fold mountains developed in resistant Palaeozoic conglomerates and sandstones (Denison Range - The Thumbs etc., Gordon Range - Tiger Range), and the great solution valleys developed on the Ordovician limestone (Vale of Rasselas, Florentine Valley). The irregular rainforest-covered area of unmetamorphosed Precambrian rocks of the Jubilee Block in the southeast forms another province.

Most of the drainage is via the Gordon River, which flows south down the Vale of Rasselas before swinging sharply west to cut across the grain of the country to the west coast. The Florentine River, which flows northeast into the Derwent, is a captured tributary of the Gordon. This capture is migrating up the flat southern part of the Vale of Rasselas, and capture of the head of the Gordon in this area is geologically imminent. Probable remnants of the former valley floor developed when the Florentine flowed into the Gordon can be seen in parts of the Florentine Valley at an altitude of about 1500 feet (Corbett, 1964). A sequence of Cainozoic gravels, sands and charcoal-bearing clays is exposed at this level on the Gordon Road near the old Needles Camp, and was probably deposited on this old valley floor.

High-level Pleistocene glacial features are preserved on a number of mountains and ranges in the area, including Wylids Craig, Denison Range, Mt. Wedge and Mt. Mueller. The features consist mainly of small cirques, moraines and moraine-dammed lakes, and in nearly all cases are restricted to the eastern side of the range crests. Although Jennings and Banks (1958) have postulated that this mountain glaciation is the only stage of the Pleistocene glacial period represented in Tasmania, evidence is accumulating which suggests that there was an

older and more extensive glaciation also. The author has previously recorded an extensive sheet of bouldery till-like material on the valley-floor just west of The Needles (Corbett, 1964) which was apparently derived from a large composite cirque between Mt. Mueller and The Needles. Other low-level deposits of this type occur just west of the Tim Shea-Needles saddle, on the Gordon Road east of Mt. Mueller, on the old South Gordon Track south of Mt. Mueller, on the Scotts Peak Road west of Mt. Eliza (an extensive till-plain here has dolerite boulders up to 10 feet long), and in the Vale of Rasselas just northwest of Battlement Hills. The deposits in the latter area appear to be at least 200 feet thick, and contain boulders of conglomerate and sandstone apparently derived from low cirque-like features on the northwesterly continuation of Battlement Hills.

Previous Literature

The general inaccessibility and harsh climate of the Adamsfield area, and the lack of economic mineral deposits apart from the osmiridium, have discouraged any major geological investigation of the area until the present study. The only important references are the report by Nye (1929) on the osmiridium field at Adamsfield, in which the general structure and the relationships of most of the major formations near Adamsfield were correctly interpreted, and the discussion and interpretation by Carey and Banks (1954) of the Lower Palaeozoic unconformities (particularly those at Adamsfield and Tim Shea). Both these reports were accompanied by sketch maps. The northern part of the Florentine Valley and Gordon Range were mapped on a regional scale by I.B. Jennings (1955), and the major part of the Florentine Valley was mapped in some detail by the author as part of an Honours project in 1963 (Corbett, 1964). Subdivision of the Ordovician limestone sequence resulted from the latter study.

The other references, although fairly numerous, are of little significance to the present study. They fall into six main groups, viz. (i) early exploration of the area (Johnston, 1888; Twelvetrees, 1908, 1909; Ward, 1909; Hills, 1921; Lewis, 1924, 1940); (ii) brief notes on the osmiridium field (Nye, 1927; Reid, 1921, 1925; Elliston,

1953; Taylor, 1955); (iii) reviews and arguments concerning the nature and relationships of the Ordovician succession (Gould, 1866; Thomas, 1945, 1948, 1960; Carey, 1947, 1953; Hills and Carey, 1949; Banks, 1957, 1961, 1962b; Williams, 1967); (iv) descriptions of Ordovician fossils from various localities (Etheridge, 1904; Chapman, 1919; Kobayashi, 1936, 1940; Teichert, 1947; Teichert and Glenister, 1953; Brown, 1948; Opik, 1951; Banks and Johnston, 1957; Thomas, 1960); (v) hypothetical suggestions concerning the general tectonics of the area (Campana et al., 1958, 1959, 1963; Scott, 1959, 1960; Carey, 1960); and (vi) unpublished reports by exploration companies of reconnaissance investigations of parts of the area (Kingsbury, 1961; Gebert, 1965, 1966; Hall, 1967). Brief descriptions of some of the Precambrian rocks near The Needles are contained in reports by Henderson (1939), Hughes (1952), Spry (1962) and Jago (1966), while brief references to the Cambrian rocks at Adamsfield are contained in reviews by Banks (1956, 1962a). The present study is based entirely on the author's own mapping, however, incorporating the earlier work on the Florentine Valley.

Acknowledgements

The continuous encouragement given by my supervisor, Mr. M.R. Banks, is gratefully acknowledged, and the many discussions with him have been invaluable to this study. The fossil identifications provided by Mr. Banks and by Dr. P.G. Quilty (formerly of this department) and Mr. J.B. Jago (Geology Department, University of Adelaide) provide the basis for some of the stratigraphy herein.

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Specimens and Thin Sections

Specimens and thin sections are housed in the Geology Department, University of Tasmania. A list of specimen numbers, with localities, rock types etc. is given in Appendix B.

CHAPTER 1

GEOLOGICAL SETTING - A BRIEF ACCOUNT OF THE TECTONICS
AND STRATIGRAPHY OF THE ADAMSFIELD AREAA. MAJOR PALAEOZOIC ELEMENTS OF WESTERN TASMANIA

The author's work in the Adamsfield area, and the recent work by other geologists in western and southwestern Tasmania, has necessitated some revision and expansion of the terminology of the major structural and palaeogeographic elements of western Tasmania. A map of some of these revised elements is shown in Fig. 1. This map has been compiled from numerous sources, including Mines Department map sheets where available, unpublished theses at the Geology Department, University of Tasmania (Groves, 1968; Gee, 1967; Rubenach, 1967; Solomon, 1964; Pike, 1964; Burns, 1963), the Geology of Tasmania (Jour. Geol. Soc. Austr., vol.9, part 2, 1962), unpublished maps and reports on southwest Tasmania by the B.H.P. Company (for which grateful acknowledgement is made), unpublished information on this area by the Hydro-Electric Commission (which is also gratefully acknowledged), and from personal communications with many people over the last four years. Only the information on the Adamsfield area is original material contributed by the present author, together with some re-interpretation and revision of the major elements as discussed below.

1. Precambrian Elements

The Precambrian elements comprise a series of structural highs, most of which were geanticlines in the Lower Palaeozoic. The major one of these is the "Older" Precambrian Tyennan Geanticline, in the core of Tasmania, and this is flanked to the west and north by a peripheral series of smaller units, viz. Cape Sorell Block, Rocky Cape Geanticline, Forth Nucleus, Asbestos Range Geanticline. The Rocky Cape Geanticline (Gee, 1967), which is the largest of these, consists mainly of unmetamorphosed "Younger" Precambrian rocks with a superimposed Cambrian basin to the west (Smithton Basin) and a strip of metamorphosed rocks (Arthur Lineament) towards the eastern side.

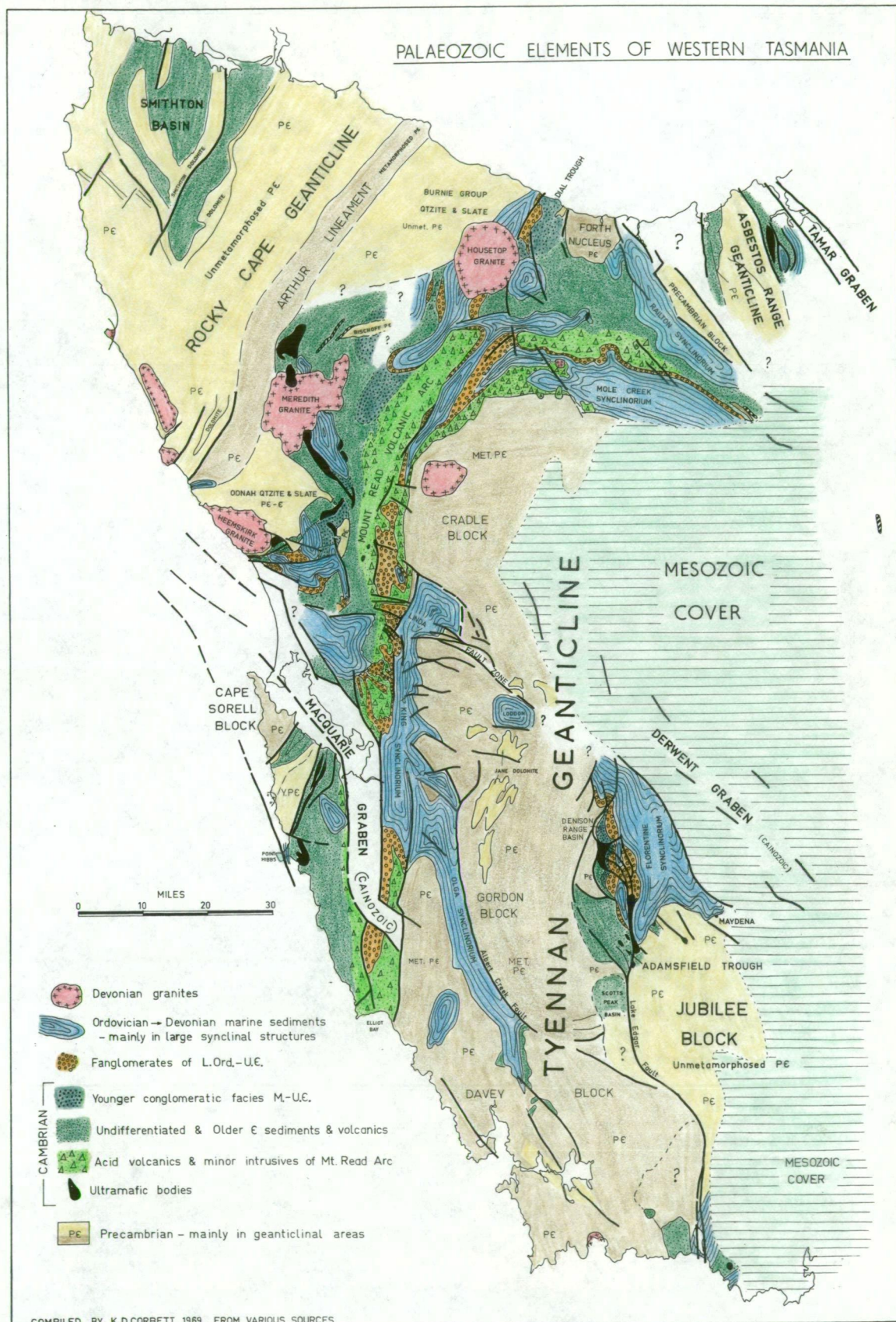


FIGURE 1

The western side of the Tyennan Geanticline abuts against or is overlapped by Lower Palaeozoic rocks and is fairly clearly defined. The position of the eastern margin, however, has always been in some doubt, partly because of overlap by the Permian-Mesozoic and partly because of lack of knowledge of the geology of the southern part. As originally proposed by Carey (1953) the "Tyennan Block" occupied most of the central part of Tasmania, including the Adamsfield area. It was named for the "Tyennan Unconformity" between Ordovician and Precambrian rocks at Tim Shea (head of Tyenna Valley), and was thought to be an area devoid of Cambrian sediments. The existence of a thick Cambrian sequence at Adamsfield was not suspected, although Nye (1929) had earlier correlated the slates and cherts just west of Adamsfield with the Cambrian "Dundas Series". The name "Tyennan Geanticline" was first used by Banks (1956), who drew the eastern margin just west of the Cambrian rocks at Adamsfield, thereby excluding the Precambrian rocks east of this and also the type area at Tyenna. A later contribution by this author (Banks, 1962a, p.128), however, shows the eastern margin back in its previous position but with a deep indentation in order to exclude Adamsfield but to take in the Tyenna area. The uncertainty regarding the eastern margin is also reflected in the contributions by Solomon (1962) and Spry (1962).

It is now recognized that the Precambrian rocks east of the Adamsfield area are mostly of the unmetamorphosed "Younger" Precambrian type, characterized by thick dolomite sequences, and probably younger than the schists and quartzites of the main part of the Tyennan Geanticline further west (e.g. Spry, 1957, 1962). Furthermore, mapping by geologists of the B.H.P. Company, Hydro-Electric Commission, and by the present author, has shown that the boundary between these two units is a major fault system extending from New River Lagoon on the south coast to the Battlement Hills area in the north (Fig.1). Part of this fault system was first noted at Lake Edgar, where there is evidence of recent displacement, by Carey (1960). A series of late Precambrian to Lower Palaeozoic basins, including the Adamsfield Trough, Scotts Peak Basin, and another in the New River Lagoon area, are developed between the two blocks, suggesting that they have been largely independent of

one another since the Precambrian. It is logical, therefore, to remove this eastern block from the Tyennan Geanticline and to give it a separate name. It is herein called the Jubilee Block, after the Jubilee Range, a large anticlinal structure south of Mt. Mueller.

The Tyennan Geanticline as presently envisaged, then, is the strip of metamorphosed "Older" Precambrian rocks extending from the south coast almost to Mole Creek, and limited to the east by the Adamsfield Trough and the fault system of which the Lake Edgar Fault is a part. The type area at Tyenna is thus excluded, but this should not invalidate a name which is firmly entrenched in Tasmanian literature. The structure definitely supplied detritus to the fringing basins during the Upper Cambrian-Lower Ordovician period, and was probably a relatively high area during most of the Palaeozoic.

There are two areas where Palaeozoic rocks extend deeply into the Tyennan Geanticline from the west, viz. the Olga Synclinorium and the northern end of the King Synclinorium (Fig. 1). Both these areas are characterized by major fault zones, viz. Albert Creek Fault, Linda Fault Zone. The Albert Creek Fault extends right through the geanticline to connect with the Lake Edgar Fault and Scotts Peak Basin. The eastward limit of the Linda Fault Zone is not yet known. These structures outline a central block within the Tyennan Geanticline which probably behaved as a separate unit during the Lower Palaeozoic, and which is herein called the Gordon Block (after the Gordon River which transects it). The segment to the north, between the Linda Fault Zone and the Mole Creek Synclinorium, appears to be a single unit, for which the name Cradle Block (after Cradle Mountain) is suggested. The name Davey Block (after Port Davey) is suggested for the segment to the south. This block includes a basin of "Younger" Precambrian rocks at Bathurst Harbour, and appears to be more complex than the other two.

2. The Cambrian Elements

The Cambrian rocks, which are predominantly of eugeosynclinal type, occur mainly in a great arcuate belt up to 30 miles wide around the western and northern margin of the Tyennan Geanticline (Fig. 1). Branches or arms of this belt radiate outwards between the fringing series of smaller Precambrian blocks (e.g. Dial Trough). The only major

break in the exposure of this belt is at the southern end of the King Synclinorium, which area may be an ancient graben structure connecting the Olga Synclinorium to the present Macquarie Graben. The equivalent of this Cambrian belt on the eastern side of the geanticline is a discontinuous series of basins, including those at New River Lagoon and Scotts Peak, and the Adamsfield Trough to be discussed below.

Two main elements have been distinguished within the west coast belt by Campana *et al.* (1963), viz. the Dundas Trough of mainly sedimentary rocks to the west, and the Mt. Read Volcanic Arc to the east. The Mount Read Arc consists mainly of acid volcanics and sub-volcanic intrusives, and appears to continue south to Elliot Bay and north to the Mole Creek area. The Dundas Trough is a complex, locally variable feature dominated by thick flysch sequences associated with mainly basic volcanics and mafic and ultramafic intrusives. Inliers of Precambrian rocks, possibly representing palaeogeographic ridges, occur at Dundas and Bischoff. Distinction can be made in places between older Cambrian sequences, characterized by greywackes and argillites (e.g. Crimson Creek Argillite), and younger (M.-U. C.) fossiliferous sequences (Dundas Group etc.). The latter are characteristically conglomeratic, and apparently much more restricted in extent than the older rocks. The ultramafic rocks separate these sequences in places, and ultramafic detritus occurs in conglomerates at the base of the Dundas Group (M.J. Rubenach, *pers. comm.*). A similar distinction is present in the Adamsfield Trough between the older Cambrian greywackes, cherts, mudstones etc., and the conglomeratic Upper Cambrian sequences, which also contain ultramafic detritus. These younger Cambrian sequences appear to reflect the increased tectonic activity associated with, and subsequent to, the emplacement of the ultramafic rocks onto the seafloor.

The Adamsfield Trough is the name given by the author to the belt of Cambrian rocks extending from the Sentinel Range north to Battlement Hills (Figs. 1,2,3). It is sandwiched between the Tyennan Geanticline (Gordon Block) and the "Younger" Precambrian Jubilee Block, and is overlapped to the east by the Ordovician rocks of the Florentine Synclinorium. It appears to have been formed by a combination of separation and sinistral shearing between the two Precambrian blocks,

and consists of an en-echelon series of tensional basins which apparently become younger to the north, as discussed below. It differs from the main west coast belt in the absence (apparently) of an acid volcanic arc, and in having an en-echelon arrangement of lithological and temporal units rather than the parallel arrangement which appears to be present in the west coast belt. Marginal strips of Precambrian from the Tyennan Geanticline have been partly incorporated into the trough, and form hook-shaped or S-shaped structures partly separating the younger basins. Ultramafic rocks were intruded along some of the major fault lines during the Middle and Upper Cambrian. The tectonics of the trough are discussed in a later section.

Campana et al. (1958) postulated the existence of an "Adamsfield Rift Valley" extending from the south coast through Adamsfield and possibly continuing as far north as the Dial Range. The structure was supposed to be a continental graben in which the Lower Ordovician conglomerates were deposited. The hypothesis was based on very sketchy information and Scott (1958) was sceptical of the existence of such a structure. The author's work has shown that the Adamsfield structure is of Cambrian and not Ordovician origin, and is essentially a eugeosynclinal trough rather than a continental graben. The conglomerates represent the final molasse-type filling of the Cambrian basins, and interfinger to the east with marine deposits. The term rift valley is not suitable, and was not used by Campana et al. (1963) in their later contribution.

3. Upper Cambrian-Lower Ordovician Elements - the Owen-type Conglomerates

A series of thick lenses of red to purple conglomerates is superimposed on the Mt. Read Volcanic Arc, and similar lenses occur in places along the Dundas Trough (e.g. Zeehan, Mt. Pearce, Valentines Peak), in the Dial Trough, and in the Adamsfield Trough (Fig. 1). The lenses along the Dundas, Dial and Adamsfield Troughs are closely associated with the younger conglomeratic sequences of the Cambrian, and appear to represent a period of more active uplift of the basin margins and deposition under mainly non-marine (alluvial fan) conditions. The relationship to the Cambrian in these areas varies from conformity to disconformity to slight unconformity.

The lenses along the Mt. Read Arc are mostly unconformable on the underlying volcanics, and are derived from the Tyennan Geanticline to the east. Campana et al. (1958, 1963) postulated the formation of a continental rift valley or graben ("Owen Graben") in which these conglomerates accumulated, but as pointed out by Banks (1962b) there is little evidence for a western fault margin or for westerly-derived material. Banks suggested that uplift of the geanticline along a series of marginal faults provided sufficient relief for conglomerates to accumulate on the sheared and tilted older volcanics. The great lateral variations in thickness, however, and the absence of conglomerates over some areas, suggests that the picture is more complicated than this. Some of the variations may be due to original topography and to formation of larger fans at some points, but some tectonic control by cross-faults also seems likely (e.g. Linda area).

4. The Ordovician to Devonian Elements

A thick sequence of marine Ordovician to Devonian sediments was deposited after the marine transgression in the Lower Ordovician, and is preserved mainly in a series of elongate synclinoria superimposed on the Cambrian rocks. Several of these are sandwiched between the Mt. Read Arc and the Tyennan Geanticline (e.g. Mole Creek, Sophia and King Synclinoria), while the others tend to overlap the younger Cambrian sequences in the Dundas Trough (e.g. Zeehan, Huskisson, Mt. Pearse). The structure near Adamsfield is herein called the Florentine Synclinorium, and is discussed below. The Olga Synclinorium and the northern end of the King Synclinorium penetrate deeply into the Tyennan Geanticline on either side of the Gordon Block.

Although the synclinoria are largely the products of the Devonian Tabberabberan Orogeny, it seems likely that some of them were areas of fairly continuous subsidence since the Ordovician, and the effect of the orogeny may have been to accentuate the original basin form into a synclinal form.

5. Tabberabberan Elements

Sedimentation ceased in the Middle Devonian with the advent of the Tabberabberan Orogeny, which was associated with the intrusion

of a number of granite bodies within the Precambrian blocks and along the Dundas Trough. The effects of the orogeny on the Palaeozoic rocks have been discussed by Solomon (1962) and earlier by Carey (1953). Apart from the large open folds paralleling the margins of the Precambrian blocks (mainly evident as the large synclinoria just discussed), a strong system of later northwest folds and faults (thrusts ?) was developed in some areas, particularly along the Mt. Read Arc. Carey (1953) interpreted these as due to dextral transcurrent movement on a major shear system ("Great Lyell Shear") along the margin of the Tyennan Geanticline, while Solomon suggested NE-SW compression. These northwest cross-folds are not developed in the Adamsfield-Florentine area.

B. TECTONICS OF THE ADAMSFIELD-FLORENTINE AREA

1. Structures of the Tyennan Geanticline

The structural geology of the Tyennan Geanticline in this area has not yet been examined in detail. Powell (1969) has recorded 3 phases of folding, all with axes trending roughly north-south, at the Middle Gordon damsite, about 10 miles west of the area mapped. In the Adamsfield area the major fold trends swing in a broad gentle arc from NE-SW near Battlement Hills to NNW-SSE near the Gordon Road, from which point there is an abrupt swing to an east-west trend on the Sentinel Range (Figs. 2,3). A similar swing in trend is evident in the Wings Lookout structure just west of Adamsfield.

Southwest of the Sentinel Range this swing in trend of the Precambrian ranges from north-south to east-west is a dominating feature, and the ends of some of the ranges appear to project into the Scotts Peak Basin, e.g. Mt. Solitary (Fig. 3) and possibly Scotts Peak itself.

2. Structures of the Jubilee Block

The structural geology of the Jubilee Block remains virtually unknown. Reconnaissance examination of outcrops and aerial photos of the northern part of the block shows a series of large folds trending about NW or NNW (Figs. 2,3). Most of the anticlines have cores of quartzite, while the synclinal areas are occupied mainly by dolomite

(e.g. "Stephens Dolomite" at Tim Shea, dolomite of the upper Weld Valley, dolomite south of Mt. Bowes) with a small Cambrian trough at Mt. Meuller. Most of the folds appear to be isolated by major faults, and there is a strong suggestion of an en-echelon arrangement. The folds plunge steeply to the NW, and have steeply-dipping limbs. The eastern limb of the Mt. Bowes structure is partly overturned.

The age of these folds is problematical. They are strongly oblique to the trend of the large Tabberabberan folds of the Florentine Synclinorium, and do not appear to be related to the Tabberabberan structures. On the other hand the NW trend is not compatible with the sinistral north-south shear system which appears to have operated during the Cambrian. Thus it is possible that the folds were formed prior to the Cambrian, but much more work remains to be done on this problem.

3. Tectonics of the Adamsfield Trough

a. General

The Adamsfield Trough is a complex feature faulted against and overlapping the Tyennan Geanticline to the west and overlapped by the Ordovician rocks of the Florentine Synclinorium to the east. Its southern margin is formed by the Sentinel Range, a strip of Precambrian quartzites projecting out from the main mass of the Tyennan Geanticline. A similar strip of Precambrian projects into the trough at Wings Lookout, and a much smaller strip at Trial Ridge (Figs. 2,3). The northern end of the trough is poorly defined because of overlap by the Ordovician rocks, but it appears to include the Battlement Hills section and probably also the thick sandstone sequence within the overlying Gordon Group in this area. The eastern margin of the trough is also poorly defined because of overlap, but for the present purposes can be taken as the Adamsfield Fault System (Fig. 3). Other Cambrian basins may be buried beneath the Florentine Synclinorium.

The shape of the trough, the arrangement of the faults, and the presence of included strips of Precambrian, suggest that the trough consists of an en-echelon series of tensional basins, perhaps formed by the action of a sinistral shear system on the margin of the Tyennan Geanticline (Fig. 3). The apparent younging of the basins to the north, as discussed below, also suggests this. The projecting

STRUCTURAL ELEMENTS OF ADAMSFIELD TROUGH AND ADJACENT AREAS

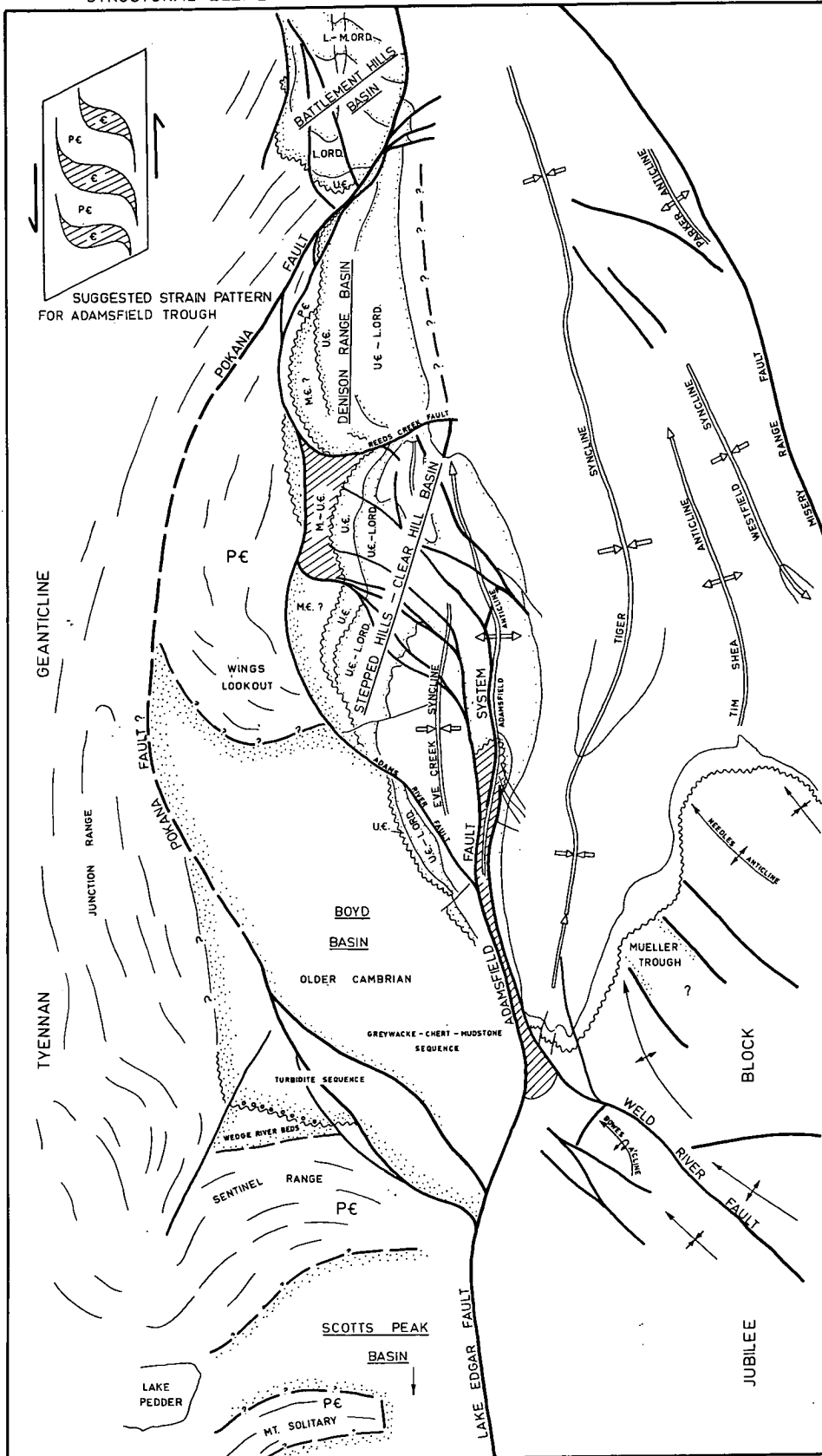


FIGURE 3

strips of Precambrian could represent marginal slices from the geanticline which were caught up in the movement and dragged around to form hook-shaped or S-shaped structures while the younger rocks were deposited partly in the tensional basins between these strips. A similar interfingering of twisted basement strips with younger basins also appears to characterize the Scotts Peak Basin to the south.

In the Adamsfield Trough the fault system appears to consist of two major fault zones, viz. the Pokana Fault to the west and the Adamsfield Fault System to the east (continuous to the south with the Lake Edgar Fault), connected by a complex series of predominantly NW-trending tensional cross-faults (Fig. 3). The Pokana Fault is exposed only at its northern and southern ends, but appears to be continuous beneath the cover of Quaternary gravels along the Pokana Valley and Denison Plain. A branch of this fault runs into the elbow at the western end of the Sentinel Range. The Adamsfield Fault System comprises several inter-connected north-south faults extending from the southern end of the Saw Back Range to the northern end of Mt. Wright. Its northerly continuation may be buried beneath the Ordovician in the Vale of Rasselas.

There is little evidence for transcurrent movement on the major faults in the area mapped (except perhaps at Battlement Hills, Fig. 2), since the rocks on either side are mainly post-Cambrian and have been affected only by the Devonian movements, which appear to have been mainly vertical. Further south, however, in the Scotts Peak area, the rocks are mainly pre-Ordovician and there is a pronounced lack of continuity of lithological features across the Lake Edgar Fault. A dolomite-limestone sequence just east of Scotts Peak appears to have been displaced northwards (sinistrally) at least 8 miles to the Mt. Bowes area, and the abrupt termination of all features against the fault strongly suggests considerable pre-Ordovician transcurrent movement.

The stratigraphy of the Cambrian rocks, which is described in the second half of this chapter, indicates that all parts of the Adamsfield Trough were not filled at the same time. Rather it appears that the filling becomes progressively younger to the north, from early

Cambrian and possibly Precambrian in the southern part to Upper Cambrian and Lower Ordovician in the northern part (Figs. 2, 3, 4). This suggests that basin-formation and basin-filling proceeded northwards with time, with relatively minor deposition of the younger rocks in the older part of the trough (prior to the Ordovician marine transgression). The distribution of the rocks, together with the arrangement of the faults and the location of the basement strips, makes it possible to subdivide the trough roughly into 4 main units, viz. the Boyd Basin in the south, the Stepped Hills-Clear Hill Basin, the Denison Range Basin, and the Battlement Hills Basin in the north (Figs. 3, 4). It must be emphasized, however, that each of these units is probably a complex of several lesser units, and that there were connections between the main units at some stages.

b. Boyd Basin

The Boyd Basin can be roughly defined as that area of mainly Cambrian rocks lying between the Precambrian strips of the Sentinel Range and Wings Lookout and approximately limited to the east by the Adams River Fault and the Adamsfield Fault System (Fig. 3). Its western and northwestern boundaries are covered by extensive Cainozoic gravel deposits, and could be either faulted or unconformable against the Precambrian. The filling of this basin, as described later, consists mainly of an unfossiliferous greywacke-chert-mudstone, with volcanics and a few mafic and ultramafic intrusives, but also includes a turbidite sequence and quartzose conglomerates in the Mt. Wedge area. The conglomerates unconformably overlie an older metamorphosed sequence of conglomerates and argillites (Wedge River Beds), which may be of Precambrian age, in the southwest corner of the basin. This unconformity appears to represent a tectonic event of some magnitude.

The structural geology of the Boyd Basin is complex and as yet poorly known, since the area is mostly covered with dense forest and scrub and has not been systematically mapped. The exposures along the Gordon River are mostly highly sheared and faulted, most of the faults trending NW. The turbidite sequence at Mt. Wedge is closely folded about axes trending east-west, but this pattern cannot be identified in the younger (?) greywacke-chert-mudstone sequence. In the central part of

the basin the chert ridges show broad arcuate trends about axes trending roughly NW to NNW, but these trends appear to be folding older folds with axes parallel to the trends of the ridges (Fig. 2).

c. Stepped Hills - Clear Hill Basin

The complex unit of Cambrian rocks and Reeds Conglomerate extending from Adamsfield to the northern end of the Stepped Hills is tentatively called the Stepped Hills-Clear Hill Basin (Fig. 3). It is limited to the west by the Precambrian of Wings Lookout, but its easterly limit is buried beneath the marine Ordovician cover. Basal conglomerates rest unconformably on the Precambrian in the Boyes River area. At Clear Hill the contact with the Precambrian appears to be faulted, and the oldest rocks in this area are a sequence of tuffaceous greywackes, siltstones, mudstones, and greywacke-conglomerates. The age of this sequence is not known, but it could be younger than the chert-greywacke sequence of the Boyd Basin. The Boyes Valley ultramafic body intrudes these rocks, and they are unconformably overlain by the Upper Cambrian Denison Group, suggesting they are probably of Middle Cambrian age or older. The Denison Group sequence thins almost to zero at Clear Hill, towards the southern end of this basin, while the Reeds Conglomerate thins markedly towards both the northern and southern margins.

The unit is cut by a series of sub-vertical NW-trending faults connecting the Adamsfield Fault System with the Pokana Fault. These faults show post-Ordovician movement and displace the top of the Reeds Conglomerate by up to several hundred feet. Displacements are mainly vertical, with south-side-down movement in the area north of the Gordon River and north-side-down in the area south of the Gordon River. A number of these faults converge on the southern end of the Boyes Valley ultramafic body, and it is possible that they are reactivated Cambrian faults.

d. Denison Range Basin

This unit is fairly clearly defined by the Reeds Creek Fault to the south and the Pokana Fault to the west (Fig. 3). Its eastern margin is not known, but is possibly a buried continuation of the Adamsfield Fault System. The filling consists of three parts, viz.

the Trial Ridge Beds, which rest unconformably on the Precambrian basement, followed unconformably by the Upper Cambrian Denison Group, followed conformably by the Reeds Conglomerate. The Reeds Conglomerate shows very pronounced thinning towards both the northern and southern margins (see Fig. 6), from over 5000 feet in the central part to only a few hundred feet at either end. The Denison Group also shows considerable thinning to the north and south. The Trial Ridge Beds thin to zero to the north, but to the south are apparently intruded by the Boyes Valley ultramafic body. The age of the Trial Ridge Beds is not known, but their relatively undeformed nature, their apparently conformable relationship to the Denison Group further north, and their lithological similarity to the Denison Group, suggest they are probably late Middle Cambrian or possibly basal Upper Cambrian.

The continuation of Denison Group-type rocks between the Denison Range and Stepped Hills suggests there was some connection between these two basins during the Upper Cambrian, and possibly before this also. There are some lithological and palaeontological differences between the two areas, however, and the precise relationship between the two units must await further detailed study. There is some suggestion of a northward migration of the deepest part of the Denison Range Basin during its filling, since the Trial Ridge Beds have a maximum thickness near the southern end, the Denison Group slightly north of this, and the Reeds Conglomerate slightly north of this again (near Bonds Craig).

A number of minor faults affect the Denison Group sequence near Reeds Peak but do not appear to continue through the overlying Reeds Conglomerate. These faults trend roughly east-west and appear to have been active during sedimentation, as discussed in Chapter 2. It seems likely that these were tensional cross-faults connecting the major north-south faults on either side of the basin, as do the faults across the Stepped Hills-Clear Hill Basin. The great thickness of the Reeds Conglomerate wedge in Denison Range Basin probably explains why these buried faults were not reactivated during the Devonian.

e. Battlement Hills Basin

This unit lies on the western side of the Pokana Fault (Fig. 3). It is cut by several NW-trending faults which connect with the Pokana Fault near the Denison Gap. The sedimentary sequence suggests that this unit did not receive sediments until late in the Upper Cambrian, and that it remained as a separate or near-separate area of subsidence until well into the Ordovician. The filling begins with a basal breccia lying unconformably on the Precambrian, followed by several hundred feet of Upper Cambrian Denison Group sandstone which thins rapidly to the west. This is followed by about 1800 feet of Reeds Conglomerate, which also thins to the northwest. The marine sandstone and siltstone sequence of the Florentine Group shows some thickening across the Pokana Fault (Fig. 2), then apparently thins again to the west.

A thick unit (approx. 1500 feet or 450 m.) of coarse quartzose sandstone and fine conglomerate occurs within the lower part of the Gordon Group limestone sequence just north of Battlement Hills, but does not appear to continue across the Pokana Fault to the east. A much thinner sandstone unit (approx. 200 feet) occurs further south in the Rasselas Valley but appears to be at a higher stratigraphic level. This thick sandstone, therefore, may represent a late stage in the filling of the Battlement Hills Basin. Although it contains trace fossils and a few poorly-preserved brachiopods, its age has not yet been established. It is probably either Lower Ordovician or basal Middle Ordovician.

f. Cambrian Orogenic Movements

The development of the Adamsfield Trough, from early Cambrian through to Lower Ordovician, appears to have been a continuous process with movements occurring at many different times, and it is difficult to separate individual orogenic phases. The tectonic activity appears to have built up to a peak in the late Middle Cambrian or early Upper Cambrian, culminating in the emplacement of the major ultramafic bodies on the seafloor. Following this, during the Upper Cambrian, there was a transition from flysch-type to molasse-type sedimentation (Denison Group) followed by the filling or "topping-up" of most of the Cambrian

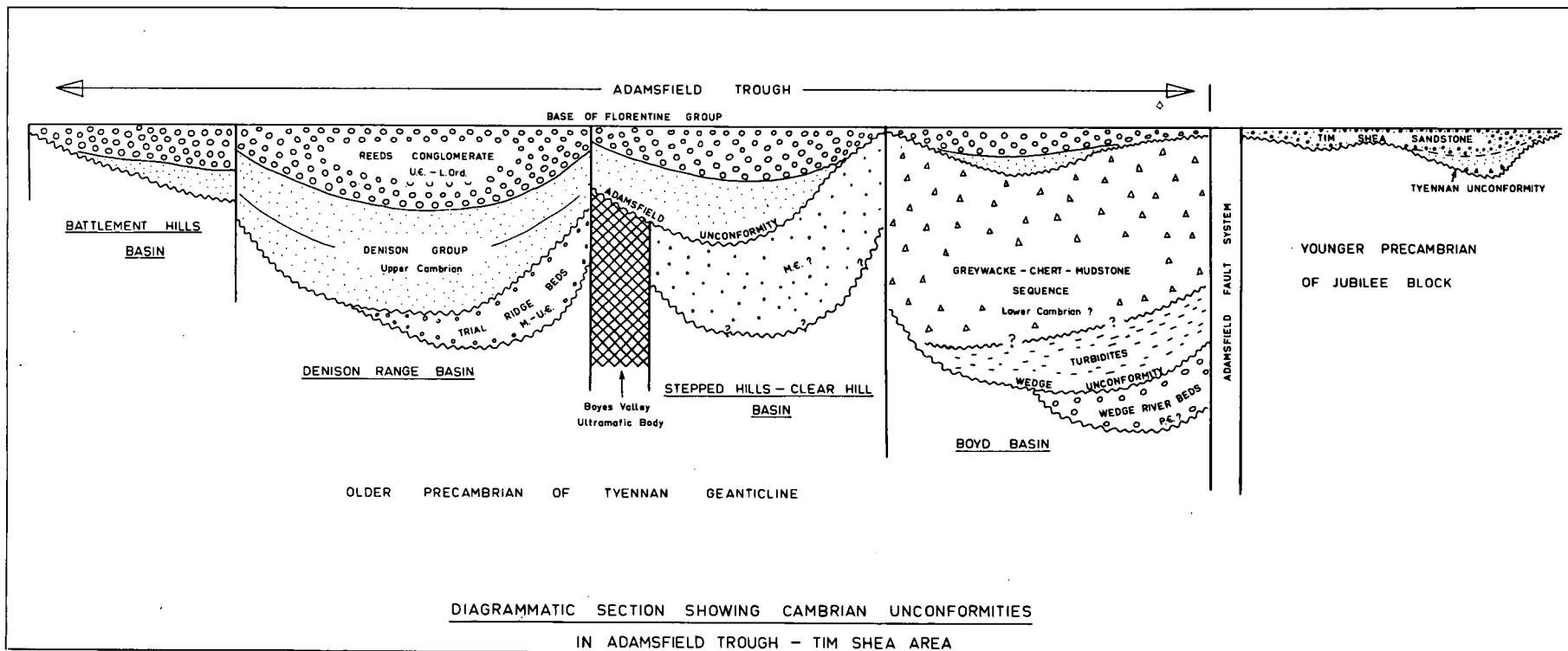


FIGURE 4

basins with thick wedges of fanglomerate (Reeds Conglomerate). Several unconformities occur within the sequence, however, which probably warrant consideration as separate tectonic episodes (Figs. 3, 4).

Unconformities between Cambrian rocks and the Precambrian schists etc. of the Tyennan Geanticline are exposed or can be inferred in a number of places along the western margin of the trough. Their significance as indicators of major movements is doubtful, however, since they appear to reflect the gradual northward migration of sedimentation with time as the successive basins were formed, and they are probably diachronous from early Cambrian to late Upper Cambrian.

The oldest unconformity within the sediments of the trough is that between the Wedge River Beds and the overlying siliceous conglomerates in the southwest corner of the Boyd Basin (Figs. 2, 3, 4). The Wedge River Beds are metamorphosed to chlorite grade and are strongly cleaved and isoclinally folded. The unconformity involves almost a right-angle change in strike. The overlying conglomerates are relatively undeformed but dip steeply northeast. They contain a few worm burrows but are otherwise unfossiliferous, and their age is unknown. They are possibly early Cambrian.

A second unconformity can perhaps be inferred between the turbidite sequence at Mt. Wedge and the greywacke-chert-mudstone sequence in the main central part of the Boyd Basin (Figs. 3, 4). The turbidites are closely folded about east-west axes, but this pattern does not appear to be present in the (? younger) greywacke-chert sequence, which shows more open folds with a predominant northwest trend. Both sequences are apparently unfossiliferous, and the contact between them is faulted, so that the age and significance of this apparent structural discordance remain doubtful.

The most extensive and perhaps the most significant unconformity is that at the base of the fossiliferous Upper Cambrian sequences (Denison Group, Adamsfield Beds, and correlates). At Adamsfield, this unconformity overlies the Adamsfield ultramafic body, and the basal conglomerates contain abundant serpentinite detritus ("Adamsfield Unconformity" of Carey and Banks, 1954). Beds just above the unconformity here contain Upper Dresbachian-Lower Franconian fossils.

A similar sequence on the Scotts Peak Road also contains serpentinite detritus. The contact above the Boyes Valley ultramafic body is not exposed but also appears to be an unconformity. Further north, this movement is expressed as a low-angle unconformity between the Denison Group and the Trial Ridge Beds (Fig. 4). Correlates of the Adamsfield Beds on the Ragged Range are unconformable on the greywacke-chert-mudstone sequence of the Boyd Basin. The movement appears to reflect the first emplacement of the ultramafics onto the seafloor, and may be of Tasmania-wide importance. The name Adamsfield Movement is suggested for this event in this area. It may be younger than the corresponding unconformity at the base of the Dundas Group (middle Middle Cambrian) on the west coast, which group also overlies serpentinite and has serpentinitic detritus at the base (M.J. Rubenach, pers. comm.).

The term "Jukesian Movement" has been applied to the postulated basement uplifts which were thought to have resulted in deposition of the Owen-type conglomerates (Upper Cambrian-Lower Ordovician) on the underlying Cambrian rocks in various parts of the state (e.g. Solomon, 1962, p.320). The type area was originally at Mt. Jukes, on the Mt. Read Volcanic Arc south of Queenstown, but the significance of this supposed unconformable contact and the length of the time-gap involved, remain in some doubt. Conformable contacts between the conglomerate and Cambrian sediments occur in a number of other areas, and it is doubtful if the Jukesian Movement involved folding. In the Adamsfield Trough the Reeds Conglomerate is mostly conformable and transitional with the underlying Upper Cambrian sediments, and appears to have been deposited in a series of basins initiated earlier in the Cambrian. Slight unconformity seems to be present at Clear Hill and possibly Adamsfield, but is probably not sufficiently pronounced to warrant distinction as a separate movement. The Tyennan Geanticline was the source area for the Upper Cambrian Denison Group (and for much of the older Cambrian sequence in the Adamsfield Trough also) as well as for the Reeds Conglomerate, so that while deposition of the conglomerate probably accompanied increased uplift of the geanticline, it did not involve a change of provenance.

The Tyennan Unconformity

The unconformity at Tim Shea, between Upper Cambrian-Lower Ordovician sandstones and breccias and the underlying "Younger" Precambrian dolomites of the Jubilee Block (Figs. 2, 4), has figured prominently in discussion of Tasmanian Palaeozoic history, and is the type area for the Tyennan Orogeny (see reviews by Carey and Banks, 1954; Solomon, 1962). As defined by Carey and Banks (1954, p.264) the Tyennan Unconformity is "the angular discordance between pre-Dundas rocks below and Junee Group above as revealed on the south-eastern slope of Tim Shea", and the Tyennan Orogeny encompasses all the "orogenic movement: occurring within the span of time represented by the Tyennan Unconformity. This span of time is at least that during which the Dundas Group was deposited." The Tyennan Orogeny has therefore been used as a general term for the summation of all the Cambrian movements prior to the Ordovician marine transgression.

The present author agrees that a general term for the whole of the Cambrian orogenic period is probably required. Some clarification of the definition of the Tyennan Orogeny appears to be required, however, since there is some evidence to suggest that one or more of the later Cambrian movements may not be represented in the type Tyennan Unconformity. The age of the Tim Shea sequence (Tim Shea Sandstone of this thesis) has not yet been definitely established, since it is mostly unfossiliferous. However, the recent discovery of Staurograptus diffusus Harris and Keble (age Lancefieldian, or very basal Ordovician) in calcareous beds of the Florentine Group several hundred feet above the top of the Tim Shea Sandstone (Dr. P.G. Quilty, pers. comm.) suggests that the bulk of the Tim Shea Formation could be Upper Cambrian. As well as this a 40-foot sequence of red siltstones which overlies the breccias in the lower part of the formation resembles similar rocks in the upper part of the Adamsfield Beds, which are Upper Cambrian.

It is possible, therefore, that the Tyennan Unconformity is about the same age as the Adamsfield Unconformity (i.e. lower Upper Cambrian), and hence any movements which occurred after this (e.g. those during Denison Group and Reeds Conglomerate time) may not be represented. While this doubt remains it would seem preferable not to

use the unconformity at Tim Shea in the definition of the Tyennan Orogeny, but to regard the latter as encompassing all the movements of the Cambrian orogenic period prior to the marine transgression in the Lower Ordovician. A precise definition is probably not possible.

4. Tabberabberan Structures

The Tabberabberan (Devonian) structures consist of a series of broad open folds making up the Florentine Synclinorium. The major one of these is the Tiger Syncline, which contains a thick core of Siluro-Devonian Eldon Group sediments. Flanking this to the west are the Adamsfield Anticline and Eve Creek Syncline (Figs. 2, 3), and to the east the Tim Shea Anticline, Westfield Syncline and Parker Anticline. For the most part the folds trend north-south or NNW-SSE, and are distinctly oblique to the earlier folds of the Jubilee Block. The majority plunge gently north or are sub-horizontal. The Adamsfield Anticline is the steepest fold, probably because of the re-intrusion of the Adamsfield ultramafic body, which intrudes through the Ordovician rocks along the Saw Back Range. The Westfield Syncline plunges gently south and has a shallow core of Eldon Group rocks at the southern end. There is no evidence of cross-folding such as characterizes the west coast structural unit and other areas (Solomon, 1962).

The Pokana Fault and the Adamsfield Fault System, and the cross-faults connecting these in the Stepped Hills area, were re-activated during the Devonian. Although there is apparently some slight sinistral displacement on the Pokana Fault at Battlement Hills, the movements on the majority of faults appear to have been mainly vertical during this orogeny. There is no evidence of large-scale transcurrent displacements of the Ordovician rocks, and hence the displacements of older rocks along the Lake Edgar Fault in the Scotts Peak area must be pre-Ordovician.

The generally regular trend of the major folds, and the absence of cross-folding or thrust-faulting, suggest that the Tabberabberan structures could have been formed by differential vertical movements of the underlying basement blocks on major north-south faults. None of the major folds can be traced into the basement rocks to the

south, and this also suggests that an overall compressional or shear system did not operate.

C. STRATIGRAPHY

1. Older Precambrian of Tyennan Geanticline

The oldest rocks in the area are the folded metasediments of the Tyennan Geanticline. These consist mainly of white quartzites and quartz-schists, forming irregular ridges and ranges, intercalated with softer phyllites, chloritic schists, graphitic schists and sericite-muscovite-schists. A pod of altered amphibolite occurs in the Gordon River about 10 miles west of the mapped area, but igneous rocks are otherwise absent. The rocks are tightly folded on several scales but for the most part show only low-grade regional metamorphism. Quartz veins are very common. Sheared conglomerates occur in several places near Battlement Hills, some of them within major fault zones. Bedding and cross-bedding can be seen in most outcrops, and ripple marks are common. The rocks have not been studied in detail. They are similar to the "Older" Precambrian low-grade metamorphic sequences described by Spry (1962) from other areas within this geanticline (e.g. Frenchmans Cap area, Mersey-Forth area).

2. Younger Precambrian of Jubilee Block

The geology of the Jubilee Block is very poorly known, and no attempt has yet been made to describe the sequences in any detail. Good outcrops are now available along the Gordon Road to The Needles and along the Scotts Peak Road, but the areas in between and to the southeast remain almost inaccessible because of the dense rainforest. Preliminary investigations by the author, including reconnaissance visits south of Mt. Mueller, to Mt. Bowes and Mt. Anne, and to the lower Weld River, indicate that the rocks for the most part are less deformed and less metamorphosed than those of the Tyennan Geanticline. There are some sequences, however, in which deformation and metamorphism approach those shown by the "Older" Precambrian, and it is possible that these represent inliers of older rocks preserved in structural highs. An area of probable Cambrian rocks, including tuffs, greywackes and lavas, was

discovered by the author at the western end of Mt. Mueller (Figs. 2, 3), and probably represents a small trough extending south into the older rocks. Other such troughs may be present elsewhere within the block. Serpentinite outcrops have been reported from just southeast of Mt. Mueller and in the Weld River by Twelvetrees (1909) and Lewis (1924).

The rocks of the Jubilee Block are characterized by great thicknesses of dolomite, as in the upper Weld Valley-Mt. Anne area, by thick siltstone sequences as on the east flank of The Needles, and by thick quartzites, as in the anticlinal zones of The Needles, Jubilee Range and Mt. Bowes. Turbidite sequences (lithic wackes, quartz wackes) occur beneath the quartzites at The Needles, and also on the Scotts Peak Road near Lake Edgar, but appear to be relatively minor components of the total sequence. Disconformities and probably unconformities occur within the sequence, and it is apparent that the stratigraphy and structure of the block are fairly complex. The thick dolomite sequence occurs at the top of the sequence at The Needles, and may also be the youngest unit in the Mt. Anne-Weld Valley area.

The section at The Needles-Tim Shea has been briefly described by Spry (1962) and by Carey and Banks (1954) under the name "Clark Group". The formations have not been properly defined, however, and the formation and group names remain informal until further work is done. The succession given by Spry (1962, p.112) was checked by the author, and is essentially correct. It consists of turbidites and siltstones at the base (exposed in the core of the Needles Anticline), followed by laminated dolomitic siltstones which are overlain by a thick sequence of cross-bedded white quartzite ("Needles Quartzite") forming the crest of the range. Above this is a sequence of siltstones, dolomites and sandstones ("Humboldt Slate and Dolomite"), followed by the massive to thin-bedded dolomite exposed on Tim Shea ("Stephens Dolomite"). The dolomite appears to occupy a synclinal structure. Spry has correlated this dolomite with other "Younger" Precambrian dolomites elsewhere in the state (e.g. Smithton Dolomite, Jane Dolomite, Fig. 1).

3. Cambrian Unfossiliferous Sequences and Igneous Rocks

a. Older Cambrian of Boyd Basin

The unfossiliferous sequence of flysch-type rocks in the Boyd Basin is older than the known fossiliferous Cambrian sequences, and is referred to as older Cambrian. It could be Precambrian in part, however. The sequence is a thick, complex one, and includes volcanics and mafic and ultramafic intrusives as well as conglomerates, turbidites, cherts, mudstones, massive greywackes etc. The lack of outcrop and the structural complications make mapping difficult and stratigraphy almost impossible, and no attempt has yet been made to establish formations. The contact with the rocks of the Jubilee Block lies in an area of intense faulting, partly exposed along the Scotts Peak Road, and its original nature is not known. For descriptive purposes, it is possible to subdivide these rocks into four units, viz. (i) the Wedge River Beds, which may be of Precambrian age (shown as such on Fig. 2); (ii) the siliceous conglomerates unconformably overlying the Wedge River Beds; (iii) the turbidite sequence just east of these basal conglomerates; (iv) the greywacke-chert-mudstone sequence, with volcanics and intrusives, occupying the main central and eastern part of the basin.

(i) Wedge River Beds

These comprise an unusual sequence of conglomerates and laminated rocks, showing chlorite-grade metamorphism and isoclinal folding, exposed on the Gordon Road just north of the Sentinel Range. The rocks strike northeasterly and dip steeply northwest. Contacts with the "Older" Precambrian of the Tyennan Geanticline are not exposed. The sequence consists essentially of massive units of boulder conglomerate alternating with zones of well-bedded to laminated sandstones and argillites. The conglomerate units are up to 30 feet (9 m) thick and vary from argillaceous (chlorite-rich) paraconglomerates or diamictites to relatively clean sandy conglomerates. Most of the boulders and pebbles are well-rounded and composed of white to pale pink quartzite typical of the Precambrian on the adjacent Sentinel Range. The strong cleavage wraps around the boulders, many of which have been flattened, drawn out and even boudinaged during deformation.

The well-bedded zones vary in thickness from about 2 feet up to about 40 feet (12 m), and consist of three main rock types, viz. laminated sandy argillites, graded sandstones and relatively well sorted sandstones and fine conglomerates. The argillites vary from greenish-grey to black in colour, and in good exposures show well-developed micro-grading in units from 2 mm to several cm thick. There appears to be a continuous series from these graded laminae to the typical graded sandstone beds. An unusual feature of the laminated sequences is the presence of numerous pebbles, most of which occur in thin, coarse-grained bands. Isolated large cobbles also occur, however, and in some cases the younger laminae appear to abut against the sides of these, such that they resemble dropstones. The graded sandstones range up to 18 inches (45 cm) thick, and commonly have a basal pebbly division.

The association of massive poorly-sorted conglomerates with varve-like argillites and graded beds could be interpreted in terms of a glacial or glacio-marine environment, and similar sequences have been interpreted in this way in the past. More recent work has shown, however, that this type of association can also be produced in a mud-flow-turbidite environment, and a glacial origin can only be confirmed if such positive criteria as striated pavements, abundant striated clasts, or unequivocal dropped megaclasts can be found. The first two of these cannot be established in the case of the Wedge River Beds because of the intense tectonic deformation. The presence of graded sandstones with pebbles suggests the operation of powerful density currents capable of depositing most of the large clasts which are found within the laminated sequences. The large cobbles within these laminites are difficult to explain by this mechanism, however, and the possibility of a glacial origin for the Wedge River Beds is considered to be an open question at this stage.

(ii) Conglomerates overlying Wedge River Beds

A sequence of about 800 feet (240 m) of quartzose conglomerates and sandstones unconformably overlies the Wedge River Beds west of Mt. Wedge. These rocks strike NW and dip NE at 60° , and form a prominent strike ridge which is intersected by the Gordon Road. Most of the conglomerates are white to grey, with some of the coarser

lenses being pale pink. The sandstones are mainly pale greenish-grey. The conglomerates are mainly pebble to cobble grade, forming beds up to 6 feet (2 m) thick, but lenses of boulder conglomerate up to 30 feet thick also occur. Clasts are mainly sub-rounded to rounded, and composed mainly of quartzite, while the matrix is predominantly sandy.

The sandstones occur as thin intercalations between conglomerate beds and also as well-bedded zones up to 100 feet (30 m) thick. Sedimentary structures include flat bedding, parallel lamination, small- to large-scale cross-bedding (sets up to 14 inches thick), current ripple marks, parting lineation, worm burrows, and pseudonodule-type deformational structures. A mixed shallow marine and fluvial environment is suggested, possibly an unstable delta or coastal plain at the margin of the basin.

(iii) Turbidite sequence near Mt. Wedge

Immediately east of the conglomerates, and probably faulted against them, is a sequence of turbidites and siltstones exposed along the Gordon Road to the east branch of the Wedge River. Similar rocks occur on the ridge south of Mt. Wedge. The sequence is faulted against the greywacke-chert-mudstone sequence which forms the main part of the basin. The turbidites are closely folded about east-west axes and are cut by numerous small faults, so that estimation of the total thickness is difficult. The beds range in thickness from less than one inch to about 4 feet (1-2 m), and the majority are clearly graded from coarse sand at the base to silt at the top. Sole marks are rarely preserved, but include flute marks, longitudinal ridges and furrows, and load casts. The upper parts of the beds show cross-lamination and deformed lamination. Parallel lamination with parting lineation forms the central part of some beds. Graded bedding is also common in the intercalated siltstones, which also appear to be of turbidity current origin.

In thin section the sandstones consist mainly of rock fragments and quartz, with 5-15% micaceous matrix and very little feldspar. The rock fragments are mainly quartzite and quartz-schist, with some sericite-schist and rare chlorite-schist. Igneous rock fragments are lacking. The rocks may be classified as lithic wackes and quartz wackes in the sense of Dott (1964), and appear to have been derived very largely from the Tyennan Geanticline.

(iv) Greywacke-chert-mudstone sequence

The main part of the Boyd Basin is occupied by a thick sequence of interbedded greywackes, cherts and mudstones, with minor lavas, tuffs and intrusives. A major ultramafic body intrudes the sequence at the Saw Back Range. Exposures are generally poor except for the sections along the Gordon Road and Scotts Peak Road, and nearly all outcrops are sheared and faulted. Deep weathering beneath the rain-forest of the Boyd Valley has converted much of the rock into red clay, and distinguishing between igneous and sedimentary rocks is difficult in many places. Only the thicker chert bands and the ultramafic intrusions form natural outcrops of any extent.

The sequence consists mainly of alternating zones, up to several hundred feet thick, of massive greywacke, banded white chert, and red argillaceous mudstone. Bedded turbidite sequences occur in a few places, and an outcrop of greenish tuffs and agglomerates is exposed on the Scotts Peak Road. Contacts between rock types are mostly faulted. The massive greywackes are sheared, argillaceous, reddish-brown to yellowish-brown rocks with little or no trace of bedding through sections up to several hundred feet thick. The only internal structures are irregular inclusions of mudstone ranging in size from small wisps up to blocks more than 10 feet across. The abundance of these, and the absence of bedding, suggest the greywackes represent large slumped masses originally consisting of interbedded turbidites and mudstones.

In thin section the greywackes are altered and sheared, with many fractured grains. Matrix is generally abundant (20-30%), and consists of chlorite and sericite with finely-divided feldspar and quartz etc. The non-matrix components consist of about 40% quartz, 25% rock fragments, 25% feldspar (variable) and 10% muscovite. The rock fragments are mostly quartzite, with lesser amounts of schistose rocks, dolomite, and igneous rocks. Secondary carbonate is abundant in specimens from near the major ultramafic body. The rocks can be classified as mainly lithic wackes (Dott, 1964), and appear to have been derived partly from the Tyennan Geanticline and partly from local igneous sources.

The cherts are well-bedded and laminated, and occur as zones up to 150 feet thick. At least 20 of these occur along the Gordon Road section, and similar bands form ridges as far north as Adams River, and in the Scotts Peak Road area to the south. Bedding is mostly steep to vertical, but some of the zones show large-scale folding. Bedding planes are commonly marked by argillaceous material, and interbedding with red argillite is common. The colour is mostly white to grey, but green, red and black bands may also be present. A thin section of the chert shows only microcrystalline quartz with abundant tiny carbonate rhombs in the darker bands. The thickness of the chert horizons, their laminated nature, their considerable lateral extent, and the apparent lack of organic material, suggest they may be primary silica-gel deposits rather than replacement bodies or organic deposits.

The mudstones are commonly closely associated with the cherts, and form zones up to 100 feet thick. They are usually deeply weathered and oxidized to reddish-brown, but fresh samples may be green to grey in colour. The majority are laminated and very fine-grained, with a fairly distinct fissility parallel to bedding. Tiny cubic holes in some samples suggest the original presence of pyrite. The absence of current structures or graded bedding, and the uniform fine-grained nature of the mudstones, suggest they are mainly of deep-water pelagic origin.

(v) Igneous rocks of the Boyd Basin

The igneous rocks, apart from the major ultramafic intrusions (which are discussed later), are very poorly exposed and mostly too deeply weathered to permit detailed examination. At least seven small igneous bodies have been located along the Gordon Road (see Appendix C map), but only one of these is at all fresh, and specimens have been collected from only three. The largest body is an altered gabbro, about 300 feet wide, which is exposed just west of the west branch of the Boyd River. A deeply-weathered porphyritic lava, about 40 feet wide, is exposed about three quarters of a mile east of the above. A thin section shows this to be an andesite, with large phenocrysts of andesine in a groundmass of fine plagioclase and granular pyroxene. A fine-grained basic rock, porphyritic in places, is poorly

exposed about half a mile west of the 25-mile peg, where it apparently encloses a large lens of mudstone. This body is decomposed to red clay, but a thin section of a fresh core shows rare phenocrysts of pyroxene and altered plagioclase in an unusual groundmass of granular pyroxene and plagioclase.

On the Scotts Peak Road the only unweathered igneous rocks are the small serpentinite and quartz-diorite bodies exposed at "Serpentine Creek". The diorite crops out in the creek bed adjacent to the serpentinite, and in thin section consists of interlocking crystals of plagioclase, quartz and amphibole cut by chloritized shear zones. Small boulders of igneous rocks derived from ? Upper Cambrian conglomerates on the ridge just west of the Scotts Peak Road include serpentinitized pyroxenite, granophyre, quartz-diorite, and an altered intermediate lava.

b. Mt. Mueller Area

A small area of flysch-type sediments, volcanics and intrusives of probable Cambrian age occurs between the western end of Mt. Mueller and the ridge of Ordovician sandstone just to the north. The existence of these rocks is known only from a single reconnaissance visit to the area by the author, and their limits have not yet been mapped. They appear to occupy a small trough extending south into the Jubilee Block. The rocks are similar to those of the Boyd Basin except for the presence of well-preserved acid tuffs. They include turbidites, siltstones, massive greywackes, red argillites, thick horizons of banded white chert, tuffs, lavas and mafic intrusives. Some of the tuffs are massive, light-coloured porphyritic rocks consisting mainly of large quartz and feldspar (andesine ?) crystals in an abundant finely recrystallized groundmass. Many of the crystals show deep embayments and have rounded or broken edges. The igneous rocks include an andesitic lava and two doleritic bodies which are probably intrusive. Aeromagnetic maps show a small anomaly in this area, possibly corresponding to an ultramafic body.

c. Clear Hill Area

A poorly exposed and poorly known sequence including sheared greywackes, siltstones, red and black mudstones, and greywacke-conglomerates, occurs on the western slopes of Clear Hill and on the south bank of the Gordon River in this area. The greywackes are green in colour, and in thin section are very sheared and chloritized, with abundant matrix. Rock fragments consist mainly of chert and fine-grained igneous rocks, with some of quartzite. Feldspar and angular quartz grains are abundant. The age of these rocks is unknown, and more work is required in this difficult area. They appear to be overlain unconformably by the conglomerates and sandstones which form the base of the Denison Group (Upper Cambrian) correlates in this area, and are apparently intruded by the Boyes Valley ultramafic body.

d. Trial Ridge Beds

A thick sequence of conglomerates, sandstones and siltstones is exposed on Trial Ridge, just west of the Denison Range, and is herein called the Trial Ridge Beds. The beds dip east at 60° on Trial Ridge itself, but to the southeast the strike swings to almost east-west and the rocks are more deformed and folded. Similar rocks occur on a ridge just west of the Boyes Valley ultramafic body, which apparently intrudes this sequence. The base of the sequence rests unconformably on the Precambrian quartzites and schists of the Tyennan Geanticline, while the upper boundary is a low-angle unconformity beneath the Denison Group. Similar rocks further north appear to be conformable with the Denison Group, however. The succession has a maximum thickness of at least 2000 feet (600 m) on Trial Ridge but thins markedly to the north. Its age is not known but it does not appear to be much older than the Denison Group (Upper Dresbachian-Lower Franconian), and could be upper Middle Cambrian or possibly Dresbachian. It represents the first phase in the filling of the Denison Range Basin.

The basal 500 feet or so consists mainly of cross-bedded conglomeratic sandstones interbedded with pebble- to boulder-grade quartzose conglomerates, with alternating zones of red and grey colour up to 60 feet thick. Clasts consist mainly of quartzite or quartz-

schist, and are mostly well rounded. This part of the sequence is probably of shallow marine and fluviatile origin.

The upper part of the sequence consists of thick units of quartzose conglomerate alternating with thicker zones of interbedded turbidites and siltstones. The conglomerates are unusual in that although they consist wholly of clasts and sand, with no fine matrix, they have a disrupted framework of pebbles "floating" in sand, and do not show any evidence of current deposition. Distinct bedding is generally absent, although irregular bands of pebbles and cobbles are common. In thin section the sandy matrix is poorly sorted and the grains have a jumbled irregular orientation. The interbedding with turbidites, and the presence further south of definite slide conglomerates containing fragments of disrupted turbidite and shale beds, suggest the conglomerate units were formed by mass movement of some kind. Further investigation of these rocks is required. The turbidites show the typical graded bedding with sharp soles showing flute marks, load casts etc. In thin section the turbidites are poorly sorted and consist mainly of quartzite and schist fragments in a micaceous matrix. Feldspar and igneous rock fragments are absent. All the rocks of the Trial Ridge Beds appear to have been derived directly from the Tyennan Geanticline.

e. The Major Ultramafic Bodies

The two major ultramafic bodies (Fig. 2) consist largely of serpentinite, and have faulted margins in most places. Several smaller serpentinite bodies occur in shear zones on the Gordon Road west of the main belt, and another is exposed on the Scotts Peak Road. Other bodies are reported from just southeast of Mt. Mueller by Twelvetrees (1909), and from the Weld River near the southern end of the Jubilee Range by Lewis (1924, p.20). Aeromagnetic maps indicate the probable presence of sub-surface ultramafic bodies along the Pokana Fault north of Trial Ridge, and of a small body in the area of Cambrian rocks near Mt. Mueller.

The Adamsfield ultramafic body is almost continuously exposed from near the Scotts Peak Road north to Adamsfield, where it

disappears under the Adamsfield Anticline at The Thumbs. The exposed width of the body varies considerably, from three quarters of a mile at Adamsfield to only a few hundred yards at the Gordon Road, where it consists of several sheared serpentinite bodies separated by wider zones of sheared sediments. At Adamsfield the belt consists of ridges of coarse pyroxenite surrounded by serpentinite. Chromite occurs abundantly as small grains within the rock and very rarely as thin cross-cutting veins. Magnetite and osmiridium are also present, and secondary deposits of the latter in Cainozoic gravels are still being worked at Adamsfield. Placer deposits also occur in Upper Cambrian conglomerates overlying the serpentinite.

Several large inclusions of sedimentary rocks, with contact metamorphic effects, occur within the serpentinite in the northeast corner of the Adamsfield body. Two of these inclusions, each about 300 feet long and 50 feet wide, consist of bedded white quartzite and form small ridges. A third body consists of serpentinitic breccia-conglomerate, apparently deposited after an early intrusion and then engulfed as the serpentinite was re-intruded. The contact metamorphic zone along the quartzite is about 60 feet (18 m) wide, and consists mainly of red jasperoid with a box-work structure of secondary quartz veins. Thin sections show several generations of secondary quartz as aggregates and veins surrounding and replacing irregular brown patches rich in haematite and fibrous blue riebeckite. A few of these patches also contain talc, and they possibly represent remnants of the original serpentinite. The contact between the jasperoid and the quartzite is irregular, with veins of jasper penetrating the quartzite. The age of the quartzite is unknown, but it could be "Younger" Precambrian.

The breccia-conglomerate body is surrounded by serpentinite and is deep bluish-grey in colour, consisting of altered fragments of talcose material in an abundant calcareous cement. Thin sections show the fragments to consist mainly of fibrous riebeckite and talc, surrounded by carbonate and secondary quartz. Similar blue rock containing abundant riebeckite is exposed in a small trench near the serpentinite contact just south of the two quartzite bodies, and is also reported from near the contact on the Gordon Road. The age of the

metamorphism is not known, but there are numerous fragments of the typical red jasperoid in the Upper Cambrian conglomerates which overlie the serpentinite at Adamsfield.

The Boyes Valley ultramafic body lies beneath thick gravel deposits, and is exposed only in stream beds. The shape of the body is fairly clearly defined by the aeromagnetic anomaly, however. The rocks exposed consist mainly of massive serpentinite, with scattered chromite grains, but no detailed study of the outcrops has yet been made. Contacts with the surrounding rocks are not exposed. The body appears to intrude and deform the Trial Ridge Beds, since these rocks are twisted into an east-west trend along the northern margin of the body and are more deformed in this area than elsewhere. The Upper Cambrian Denison Group rocks, on the other hand, appear to be younger than the main intrusion and to overlie it, and the Trial Ridge Beds, unconformably.

Serpentinite detritus occurs abundantly in conglomerates of possible upper Middle Cambrian or Upper Cambrian age on the Scotts Peak Road, as discussed below, and in Upper Cambrian conglomerates at Adamsfield. It is apparent, therefore, that the bodies were exposed on the seafloor and sufficiently uplifted for erosion to occur by early Upper Cambrian time. The fact that much of the Cambrian conglomeratic facies was formed just prior to or just after this period suggests that the emplacement of the ultramafics was accompanied by a period of increased tectonic activity. The intrusions must have continued intermittently throughout the Upper Cambrian and possibly into the Lower Ordovician, as evidenced by the presence of serpentinite detritus in conglomerates throughout the Adamsfield Beds. Further re-intrusion of considerable magnitude must have occurred during the Devonian Orogeny to force the ultramafics up through the mantle of Ordovician sediments along the Adamsfield Anticline.

4. Cambrian Fossiliferous Sequences

Fossiliferous Cambrian sequences occur in three main areas, viz. (i) a small area on the Scotts Peak Road; (ii) above the serpentinite at Adamsfield (Adamsfield Beds); and (iii) a strip along the Denison Range and Stepped Hills (Denison Group). These sequences all appear to

be younger than the majority of those just described, and are characteristically rich in conglomerates.

a. Scotts Peak Road Sequence

An unusual sequence of about 1000 feet (300 m) of dolomitic sandstones, red and green mudstones, serpentinitic conglomerates and turbidites is exposed on the Scotts Peak Road at the first major saddle, about two miles south of the Gordon Road. The sequence is sheared and folded in places, but appears to be essentially a synclinal structure with a faulted eastern contact against a greywacke-chert-mudstone sequence similar to the older Cambrian of the Boyd Basin. The base of the sequence on the western side is a coarse breccia-conglomerate, containing blocks of argillite up to 10 feet long, unconformably overlying red argillites. Above this are dolomitic sandstones and conglomerates, with interbedded mudstones, and these are overlain by a thick sequence of turbidites with some intercalated serpentinitic conglomerates. The conglomerates contain pebbles of other igneous rocks, including dolerite, granophyre, diorite and andesite, as well as rounded fragments of serpentinite up to 5 inches (13 cm) long.

Fossils have been found in a single thin siltstone band within the turbidites just south of a small road-metal quarry, and have been identified by Dr. P.G. Quilty. The small fragmentary collection includes the dendroids Mastigograptus serialis (formerly Archaeolafoea serialis), M. simplex, and Archaeolafoea monegettae, an inarticulate brachiopod (? Palaeobolus sp.), and an indeterminable arthropod. The dendroid association occurs in Middle Cambrian rocks in Victoria and western Tasmania, but individual ranges include Upper Cambrian.

Several other small areas (fault blocks ?) of similar rocks occur just southwest of the above sequence. One of these is exposed on the Scotts Peak Road and consists of conglomerates with dolomite fragments at the base, followed by turbidites and siltstones. On the ridge west of the road there are sandstones and conglomerates containing chert detritus and some igneous rock fragments, apparently derived from older Cambrian rocks to the west.

b. Adamsfield Beds

The sequence of conglomerates, sandstones, siltstones and impure limestones unconformably overlying the serpentinite at Adamsfield, and overlain by quartzose conglomerates of the Reeds Conglomerate, is herein called the Adamsfield Beds. The sequence is exposed in excavations along the serpentinite contact at the head of Main Creek on the eastern side of the ultramafic body, and in small reservoirs and scrapes up the slope to the east of this. The total thickness is of the order of 1000 feet (300 m).

The contact with the serpentinite is faulted in most places, but the presence of a serpentinitic conglomerate up to 20 feet thick along the contact indicates an original unconformity. The conglomerate consists wholly of serpentinite detritus, including abundant chromite and some osmiridium, and is overlain by interbedded calcareous sandstones and siltstones with some impure nodular limestones. The calcareous rocks contain abundant serpentinite detritus as well as a rich marine fauna of brachiopods, trilobites, gastropods etc., and are about 200 feet thick. Above these, but poorly exposed, are laminated and cross-laminated siltstones and fine sandstones, with some intercalated thick beds of red serpentinitic conglomerate.

The upper half of the Adamsfield Beds is formed by an interesting sequence of thin-bedded quartzose sandstones, siltstones, and pebble conglomerates containing abundant worm tubes as well as gastropods, inarticulate brachiopods and very rare trilobites. These rocks are exposed in two small reservoirs towards the top of the ridge, and along the tracks leading to these. They are mostly light-coloured, but red siltstones and fine sandstones occur at the top of the reservoir sections. Chromite is common throughout the sequence, being concentrated along many lamination planes and causing green staining of the outcrops. Inter-lamination of fine sandstone and siltstone is common, and other sedimentary structures include current ripple marks, cut-and-fill structures, biogenic reworking, flat lamination with parting lineation, low-angle cross-bedding, load casts, shale pellets, and pseudonodule-type deformational structures. This part of the sequence is interpreted as being mainly of inter-tidal and near-shore origin.

The reservoir exposures mostly dip steeply east but are folded along axes plunging gently northwest. The east limbs of the anticlinal folds are shallow-dipping and overturned, suggesting the folds are "crinkles" on the steep east limb of a larger anticlinal structure (Adamsfield Anticline) formed partly by the intrusion of the serpentinite. Overlying the reservoir sequence is a series of boulder conglomerate beds interbedded with bioturbated sandstone and cross-bedded conglomeratic sandstone. This sequence is about 150 feet (45 m) thick, and is overlain by a thick sequence of red sandstones and fine conglomerates exposed across the top of the ridge. The boulder conglomerates possibly represent the base of the Reeds Conglomerate, which is very thin in this area, and there could be a disconformity or low-angle unconformity between these beds and the underlying folded reservoir sequence. The conglomerates contain abundant chromite and many large fragments of red jasperoid, suggesting renewed uplift of the ultramafic body.

A collection of brachiopods and trilobites from near the base of the Adamsfield Beds was examined by A.A. Opik (in Banks, 1962a, p.137), who suggested an Upper Dresbachian or Lower Franconian age. Mr. M.R. Banks has tentatively identified the gastropod Kobayashiella and the trilobite ? Dellea from the upper part of the sequence, again indicating an Upper Cambrian age.

Correlates of the Adamsfield Beds

The Upper Cambrian Denison Group rocks of the Denison Range-Clear Hill area must be largely equivalent to the Adamsfield Beds. The upper part of this group (Great Dome Sandstone) has many features in common with the upper part of the Adamsfield Beds, and was deposited under similar paralic conditions.

A 600-foot thick sequence of chert-rich conglomerates, quartzose sandstones and micaceous siltstones underlies the Reeds Conglomerate on the Ragged Range southwest of Adamsfield, and is probably partly equivalent to the Adamsfield Beds. This sequence rests unconformably on the older Cambrian cherts, greywackes and mudstones of the Boyd Basin. The abundance of worm tubes on some horizons, and the presence of current ripple marks, cross-lamination and flat lamination

suggest a near-shore environment for most of this sequence. Red siltstones and fine sandstones also occur in places. The sequence thins rapidly to the north and south.

A serpentinitic chert-rich conglomerate is faulted against the eastern margin of the serpentinite body on the Gordon Road at the southern end of the Saw Back Range, and is possibly equivalent to the serpentinitic conglomerate at the base of the Adamsfield Beds.

At Tim Shea, the Tim Shea Sandstone has about 40 feet of red siltstone, with numerous small worm tubes, near its base, underlain by a locally-derived dolomitic breccia up to 200 feet thick. Similar red siltstones occur in the Ragged Range and Adamsfield sequence, as just noted. The formation also contains chromite in many places, and is partly marine and partly non-marine. The recent identification of Staurograptus diffisus from the Florentine Group beds well above this formation suggests that the bulk of the formation could be Upper Cambrian. Thus the lower part of the Tim Shea Sandstone may be laterally equivalent to the Adamsfield Beds.

The Adamsfield Beds may also be equivalent to the lower part of the Reeds Conglomerate in some areas. This formation thickens from a few hundred feet at Adamsfield to more than 2000 feet (600 m) on The Thumbs just to the north (Fig. 6). The lower part of the formation in that area consists mostly of pink sandstones, but includes an horizon of grey sandstone with abundant worm burrows, suggesting a brief incursion of marine conditions. Lateral interfingering with the Adamsfield Beds seems quite probable.

c. Denison Group

The belt of Upper Cambrian sandstones, siltstones and conglomerates which underlies the Reeds Conglomerate from Battlement Hills south to Clear Hill is herein called the Denison Group. In the type area on the Denison Range it consists of two formations, viz. the Singing Creek Siltstone at the base, and the Great Dome Sandstone. The sedimentology and detailed stratigraphy of these formations forms the main part of this thesis. The group has a maximum thickness of about 4300 feet (1300 m) on the Denison Range, thinning to practically zero at Battlement Hills and Clear Hill. Significant thinning also appears

GEOLOGY OF CENTRAL AND NORTHERN DENISON RANGE

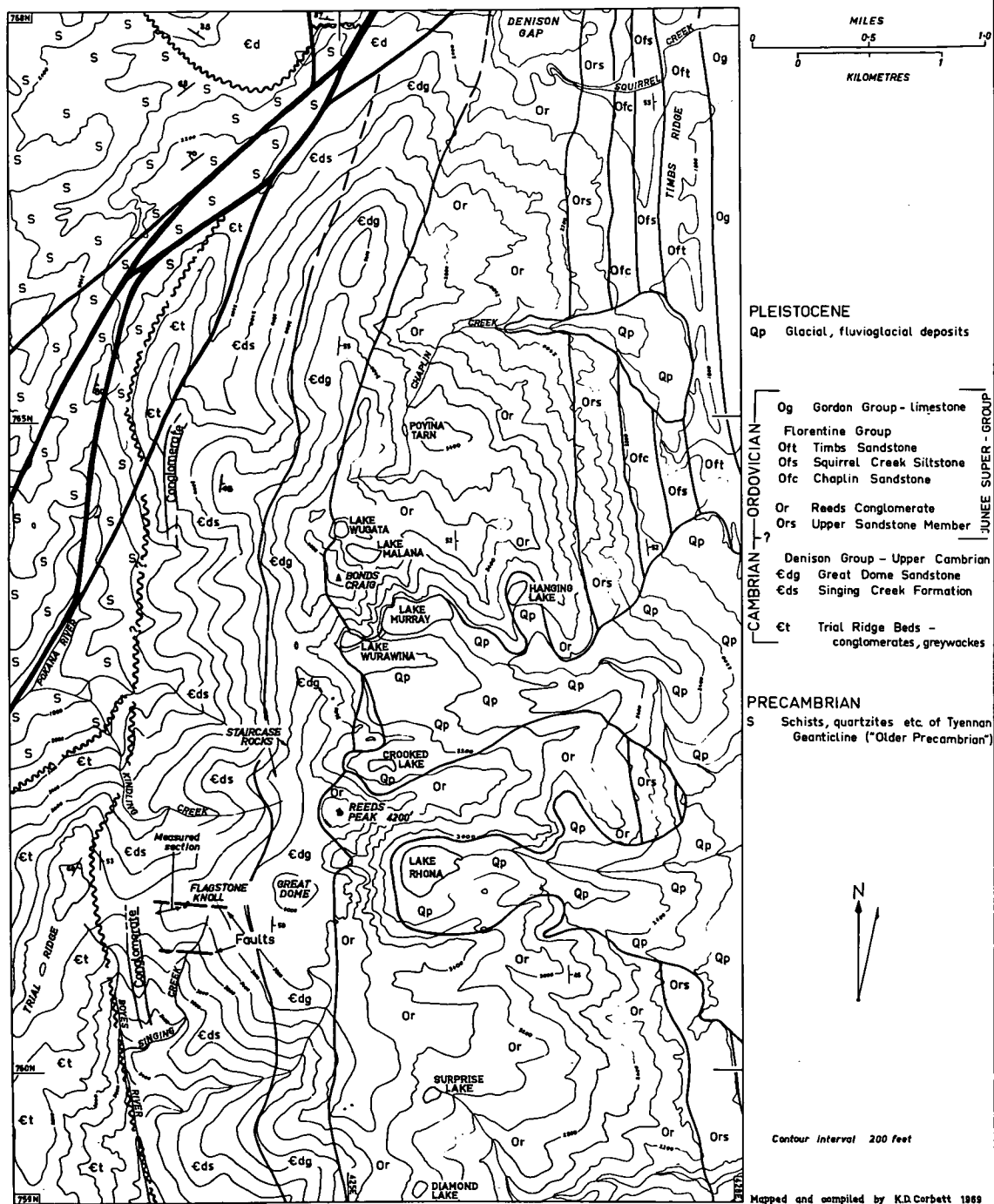


FIGURE 5

to occur in the area between the Denison Range and Stepped Hills, and the fault zone (Reeds Creek Fault) through this area probably formed the southern limit of the Denison Range Basin during most of the Upper Cambrian. Palaeontological dating indicates an Upper Cambrian age for most of the group on the Denison Range. Definitions of the two formations are given below.

(i) Singing Creek Siltstone

Derivation: Singing Creek (7604N, 4235E), a headwater tributary of the Boyes River on the western side of the Denison Range (Fig. 5).

Type section: Flagstone Knoll, Singing Creek and adjacent slopes of Great Dome; basal conglomerates exposed on northern slope of Trial Ridge (Fig. 5).

Lithology: Laminated grey siltstone-fine sandstone forms the "matrix" of the formation, and occurs as sequences up to several hundred feet thick as well as intercalations between sandstone beds. Very quartzose turbidites (pure quartz wackes in sense of Dott, 1964) occur as single beds within the siltstone, but more commonly as composite sequences forming large lenses up to at least 80 feet thick. Pebble- to boulder-grade quartzose conglomerates, many of them showing graded bedding, occur in some of these lenses. Slump sheets consisting of intermixed sandstone and siltstone are also an important component. The base of the formation is formed by a breccia-conglomerate, up to 15 feet thick, derived mainly from the underlying Trial Ridge Beds. Details of the stratigraphy and sedimentology are given in Chapter 2.

Thickness: Approximately 2600 feet (780 m).

Palaeontology and age: The siltstones contain a fairly abundant marine fauna of trilobites and brachiopods, with rare gastropods and numerous trace fossils (mainly worm burrows, some probable trilobite tracks). Some of the brachiopods and trilobites have been tentatively identified by Mr. M.R. Banks, and a detailed examination of the trilobites is being undertaken by Mr. J.B. Jago as part of a Ph.D. thesis at the Geology Department, University of Adelaide. The trilobites include Pseudagnostus cf. sentosus (M.R.B.), Pseudagnostus cf. communis (M.R.B.),

? Micragnostus sp., Proceratopyge sp., Dellea cf. suada. Inarticulate brachiopods are abundant, and include Lingulella cf. perattenuata, ?Pseudodicellomus sp., ? Palaeobolus sp. Articulate brachiopods include Billingsella cf. corrugata, Billingsella sp. nov., ? Jamesella sp.

Mr. Banks (pers. comm.) suggests a Lower Franconian age (probably Elvinia-zone), while Jago (pers. comm.) suggests an Elvinia-zone or Conaspis-zone age.

Relationships: Rests with low-angle unconformity on Trial Ridge Beds of unknown age (M.C.?); passes upwards transitionally and conformably into Great Dome Sandstone; probably equivalent to lower part of Adamsfield Beds.

Palaeogeographic interpretation: Submarine fan complex formed at foot of fault-controlled slope around western margin of Denison Range Basin. The lenses of sandstone and conglomerate are interpreted as mainly channel-fillings on fan surface, and are too laterally-impersistent to warrant member status.

(ii) Great Dome Sandstone

Derivation: Great Dome (7615N, 4246E), a prominent dome-like feature on the crest of the Denison Range near Reeds Peak (Fig. 5).

Type section: Staircase Rocks area, immediately west of Reeds Peak (Fig. 5).

Lithology: Lower half consists of units of thick-bedded coarse sandstone, showing large-scale cross-bedding, alternating with zones of thin-bedded (flaggy) sandstone and siltstone, and is referred to as the Cyclic Facies. Detailed descriptions of the stratigraphy and sedimentology of this part are given in Chapter 3. Upper half of the formation consists mainly of cross-bedded coarse sandstone and fine conglomerate, with only minor intercalations of flaggy sandstone and siltstone. Colour mostly grey, but the upper part becomes red in the area north of Bonds Craig.

Thickness: Approximately 1700 feet (510 m).

Palaeontology and age: Poorly fossiliferous except for trace fossils. The thin-bedded zones contain gastropods and inarticulates in places. Worm burrows and intensive biogenic reworking are characteristic of the flaggy sandstones in all areas. The gastropod has been tentatively identified as Kobayashiella sp. by Mr. M.R. Banks (pers. comm.), suggesting an Upper Cambrian age.

Relationships: Transitional downwards into Singing Creek Siltstone via 50-100 feet of transition beds consisting of interbedded sandstone and siltstone; also transitional upwards into Reeds Conglomerate, through several hundred feet of fine conglomerate and conglomeratic sandstone, and probably interfingers with that formation; probably equivalent to upper part of Adamsfield Beds.

Palaeogeographic interpretation: Fluvio-marine deltaic sequence, passing up into floodplain and alluvial fan facies; cycles of Cyclic Facies formed by shallow marine incursions over areas temporarily abandoned by the distributary system - see Chapter 3.

(iii) Denison Group in other areas

At Battlement Hills the Palaeozoic sequence rests with marked angular unconformity on the Precambrian quartzites and schists of the Tyennan Geanticline. This unconformity was first noted by Twelvetrees (1908). A coarse, locally-derived breccia overlies this unconformity in most areas. The Denison Group is represented in the eastern part by about 600 feet (180 m) of cross-bedded quartzose sandstones, with some intercalated bioturbated flaggy sandstones, overlying the basal breccia. This sequence thins rapidly to the west, to be replaced by the Reeds Conglomerate. The facies is similar to the Great Dome Sandstone, and there is no equivalent of the Singing Creek Siltstone in this area.

Rocks correlated with the Denison Group occur beneath the Reeds Conglomerate along the Stepped Hills. The sequence here is poorly exposed but is of the order of 2000 feet (600 m) thick. The lower part includes siltstones, turbidites and conglomerates similar to those of the Singing Creek Siltstone. Trilobites, articulate and inarticulate brachiopods, gastropods, rare bivalves, and trace fossils occur in these

lower beds, but have not yet been systematically collected or identified. Mr. M.R. Banks has tentatively identified a new species of Billingsella, and some new genera of trilobites and brachiopods appear to be present. The fauna is rather different from that of the Singing Creek Siltstone. Mr. Banks suggests an Upper Cambrian age. The upper part of the Stepped Hills sequence consists mainly of cross-bedded sandstones with intercalated flaggy sandstones and siltstones, similar in facies to the Great Dome Sandstone. Rare inarticulates and gastropods occur within this part, which includes a few horizons of glauconitic sandstone.

A poorly exposed sequence underlies the Reeds Conglomerate at Clear Hill and can probably be correlated with the Denison Group. The base of the sequence, as exposed on the south bank of the Gordon River, consists of coarse quartzose conglomerates with interbedded sandstones, and appears to unconformably overlie the sheared greywackes and mudstones etc. just to the west. The middle part of the sequence is not exposed. The upper part consists mainly of bioturbated flaggy sandstones with some intercalated thick-bedded sandstones, and is fairly similar in facies to the Great Dome Sandstone. The maximum thickness of the sequence is of the order of 1000 feet (300 m), but it appears to thin almost to zero to the south. The contact with the Reeds Conglomerate is poorly exposed, but appears to be disconformable or unconformable near the crest of Clear Hill, where the sandstones strike obliquely under the conglomerate.

5. Ordovician Stratigraphy

a. General

The Upper Cambrian sequences of the Adamsfield Trough are overlain by thick conglomerates of the Reeds Conglomerate formation, which is in turn overlain by fossiliferous Lower Ordovician marine sandstones and siltstones. These pass upwards into the thick Ordovician limestone sequence which occupies the bulk of the Florentine Synclinorium. The age of the Reeds Conglomerate is in some doubt, and it may in fact be mostly Upper Cambrian. It has become standard practice, however, to group these conglomerates (Owen Conglomerate and correlates) with the overlying fossiliferous Ordovician sequences when discussing Palaeozoic

stratigraphy (e.g. Banks, 1962b), although their palaeogeographic and tectonic affinities appear to be with the Cambrian rather than the Ordovician.

The author's mapping in the Florentine Synclinorium has necessitated a considerable expansion and revision of the stratigraphic nomenclature of the Ordovician, including the establishment of a number of new formations and groups. However, this work is not immediately relevant to the main study of this thesis, and it is therefore included only as an appendix (Appendix A). Suffice it to say at present that the sequence has been subdivided into three main units, viz. the Reeds Conglomerate (and its lateral equivalent the Tim Shea Sandstone) at the base, followed by the Florentine Group (marine sandstones and siltstones of Lower Ordovician age), followed by the Gordon Group (mainly limestones, Lower to Upper Ordovician age). These three units make up the June Super-Group. Definitions of the various formations and groups are given in Appendix A. Only the Reeds Conglomerate and Tim Shea Sandstone are discussed here because of their relationship to the Upper Cambrian sequences.

b. Reeds Conglomerate (see Appendix A for definition)

The Reeds Conglomerate is the name given to the mountain-forming unit of siliceous conglomerate stretching from the Battlement Hills to the southern Saw Back Range (Fig. 2; Plate 1A). The formation varies greatly in thickness, reaching a maximum of just over 5000 feet (1500 m) near Reeds Peak (Fig. 6). It occurs essentially as four great wedges, one centred on Battlement Hills, one on the Denison Range, one at Stepped Hills-Clear Hill, and one at the Ragged Range-Saw Back Range, corresponding to the four main basins of the Adamsfield Trough (Figs. 3, 4).

The contact between the conglomerates and the underlying Upper Cambrian sediments is conformable and transitional in most areas, but may be unconformable at Clear Hill. The upper contact with the marine Florentine Group is transitional and probably interfingering. Intercalations of marine sandstone with abundant worm burrows occur in the middle part of the conglomerate sequence at Clear Hill and The Thumbs. The upper part of the formation in most areas is formed by a

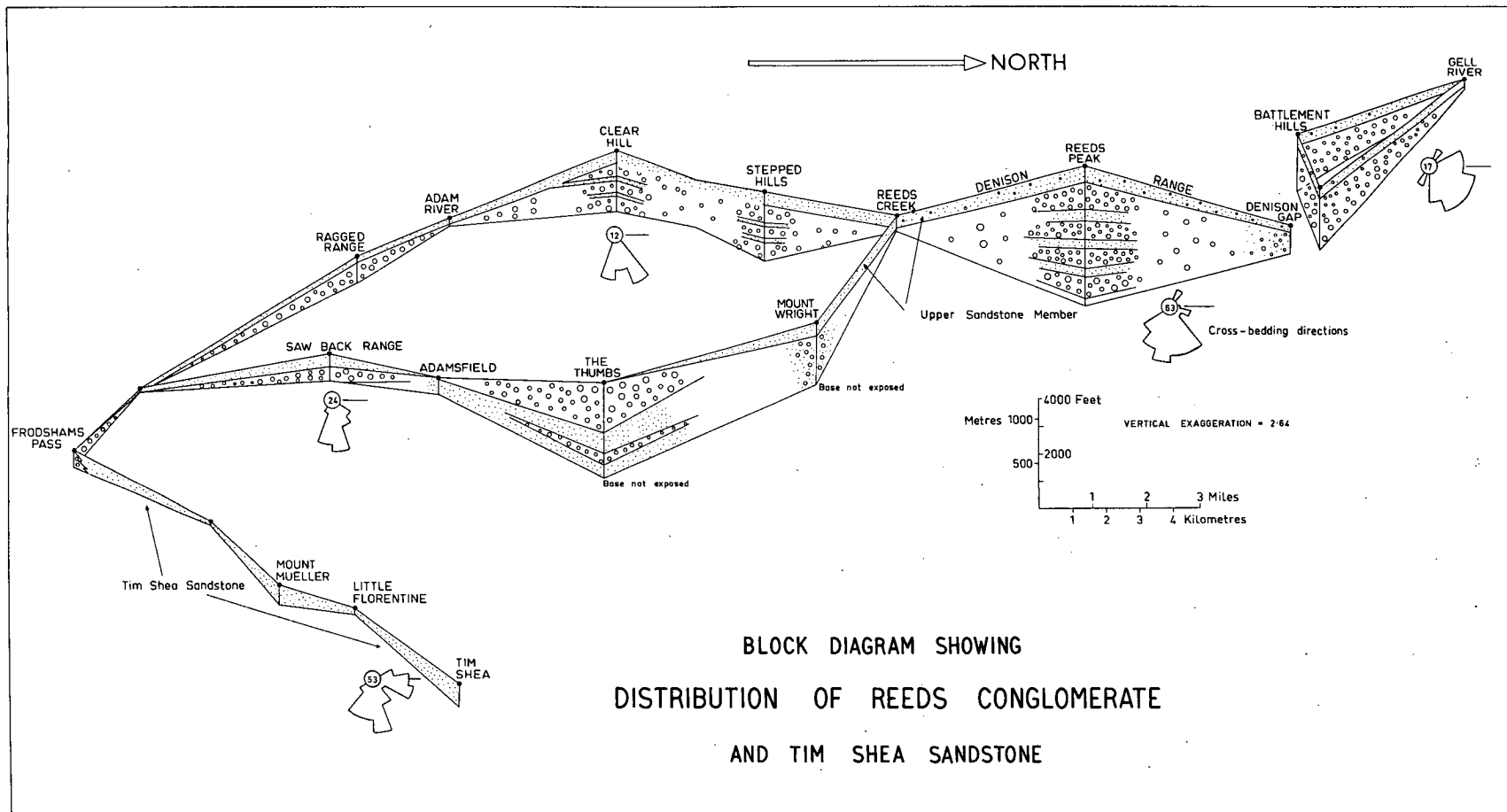
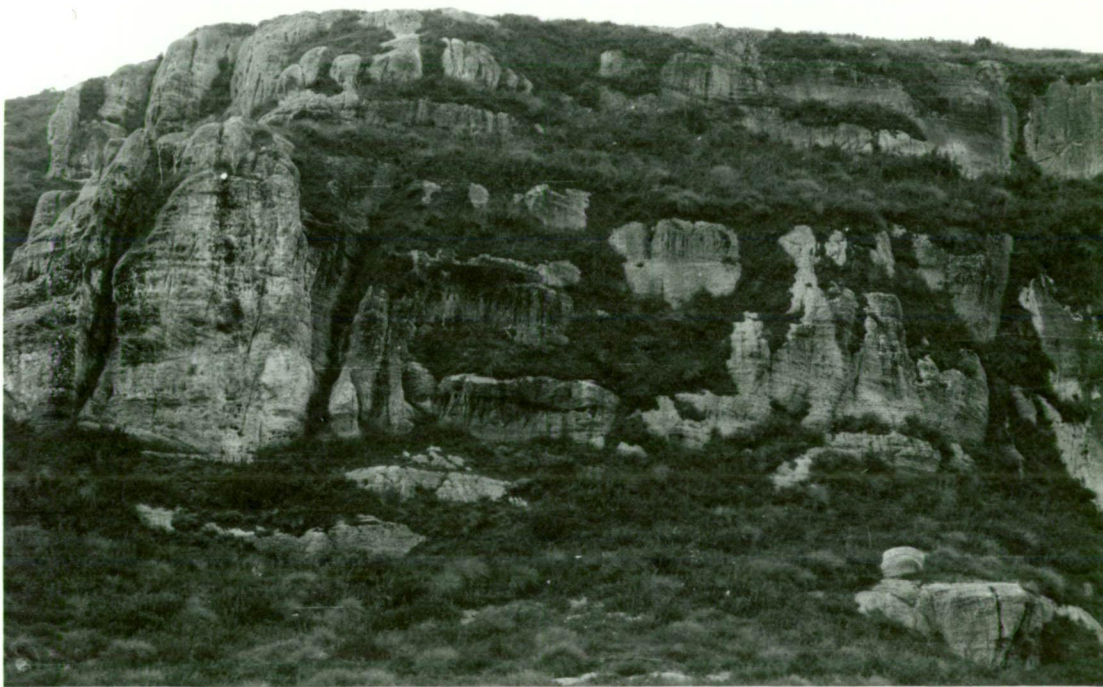


FIGURE 6



A. Reeds Conglomerate on western slope of Stepped Hills. Thick-bedded coarse conglomerates form bluffs, while sandstones and fine conglomerates form shelves. Upper Sandstone Member on crest of range.



B. Upper Sandstone Member of Reeds Conglomerate at Battlement Hills. Mostly cross-bedded conglomeratic sandstone. Tendency towards columnar weathering is typical.

sandstone or conglomeratic sandstone unit, up to several hundred feet thick, called the Upper Sandstone Member (Plate 1B; Fig. 6). This unit usually shows abundant trough cross-bedding, and varies from red to grey in colour. The grey horizons show flat bedding and contain abundant worm burrows in places, suggesting deposition under shallow marine conditions.

The bulk of the formation is red to brown to purplish in colour, and consists of pebble- to boulder-grade siliceous orthoconglomerates, with an abundant sandy matrix, interbedded with conglomeratic sandstone (Plate 2). Siltstones and finer-grained rocks are absent or extremely rare. Within the major wedges there appears to be a second-order arrangement of large lenses of conglomerate, usually several hundred feet thick, separated by sandstone. Third-order lensing on the scale of beds or groups of beds can be seen in most outcrops. The sandstones show abundant large-scale trough cross-bedding, while the finer conglomerates are characterized by abundant scour-and-fill structures, rapid inter-lensing of sandstone and conglomerate (Plate 2B), rare large-scale tabular cross-bedding, and fairly numerous large channel structures. The coarse (cobble to boulder) conglomerates tend to be thick-bedded to massive (Plates 1A, 2A), forming semi-homogeneous units up to 20 feet thick. Large channel structures and large-scale heterogeneous cross-bedding occur within the coarse conglomerates in places. Imbrication of tabular clasts is apparent in a few outcrops (e.g. Plate 2A). The red colouration is due partly to limonite-coating on the clasts and partly to interstitial limonite.

The majority of clasts consist of either quartzite or quartz-schist, and the only other common types are vein quartz, chert and quartz sandstone. Fragments of chlorite-schist, sericite-schist, phyllite, shale and rare talc occur in the conglomerate at the southern end of the Saw Back Range. Igneous rock fragments are apparently absent, as is feldspar. Most of the clasts are well rounded to very well rounded (Plate 2), while the sand-grade material, which is also predominantly siliceous, is mostly sub-angular to sub-rounded. Sphericity is moderate to high generally, although some of the more schistose clasts tend to be tabular. The great bulk of the sediment appears to have been



A. Typical outcrop of Reeds Conglomerate at Clear Hill. Massive coarse conglomerate above, with inter-lensing fine conglomerate and sandstone below. Note rounding and slight imbrication.



B. Reeds Conglomerate, northern Denison Range. Typical fine conglomerate showing scour structures, inter-lensing with sandstone, rapid changes of grain size and pebble content.

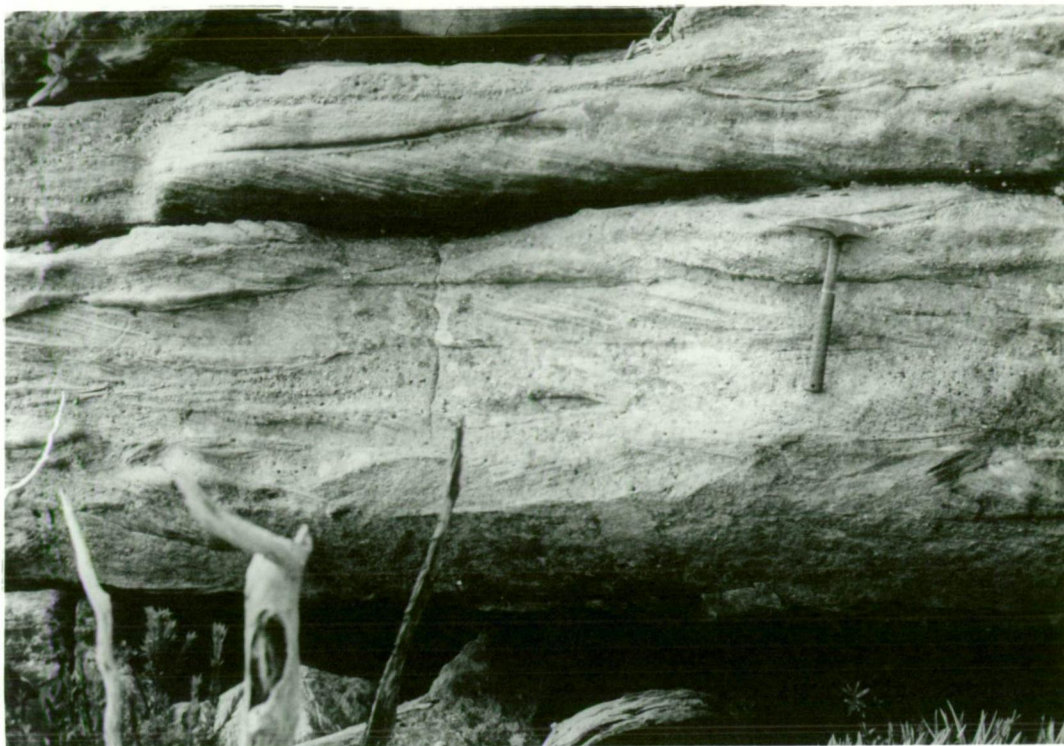
derived from the Precambrian rocks of the adjacent Tyennan Geanticline, with a small contribution (mostly chert) from the Cambrian rocks in the southern part of the area. Palaeocurrent evidence to date also indicates derivation from the Precambrian to the west and southwest (Fig. 6).

The prevailing red colour of the conglomerates, the abundance of cross-bedding and channel structures, the pronounced lateral variability of the sequence, the rounding and imbrication of the clasts, the bimodal nature of the conglomerates, the absence of very fine-grained material, and the absence of fauna except for the worm burrows in the rare marine intercalations, indicate deposition by powerful streams under non-marine conditions. The features coincide with those of modern alluvial fan complexes on which deposition is mainly by shallow braided streams (e.g. Denny, 1965; Bull, 1963; McKee, 1957; Blissenbach, 1954; Gregory, 1915; Trowbridge, 1911), and with ancient deposits interpreted as fan conglomerates of this type (e.g. Bluck, 1965; Allen, 1965d; Potter, 1955; Krynine, 1950). There is no evidence to support their interpretation as littoral deposits (Carey and Banks, 1954), although interfingering with marine sandstones occurs in places. The formation, and its equivalents in western and north-western Tasmania (Owen Conglomerate etc.) would appear to be one of the most extensive and best developed examples of alluvial fan deposition in the geological record.

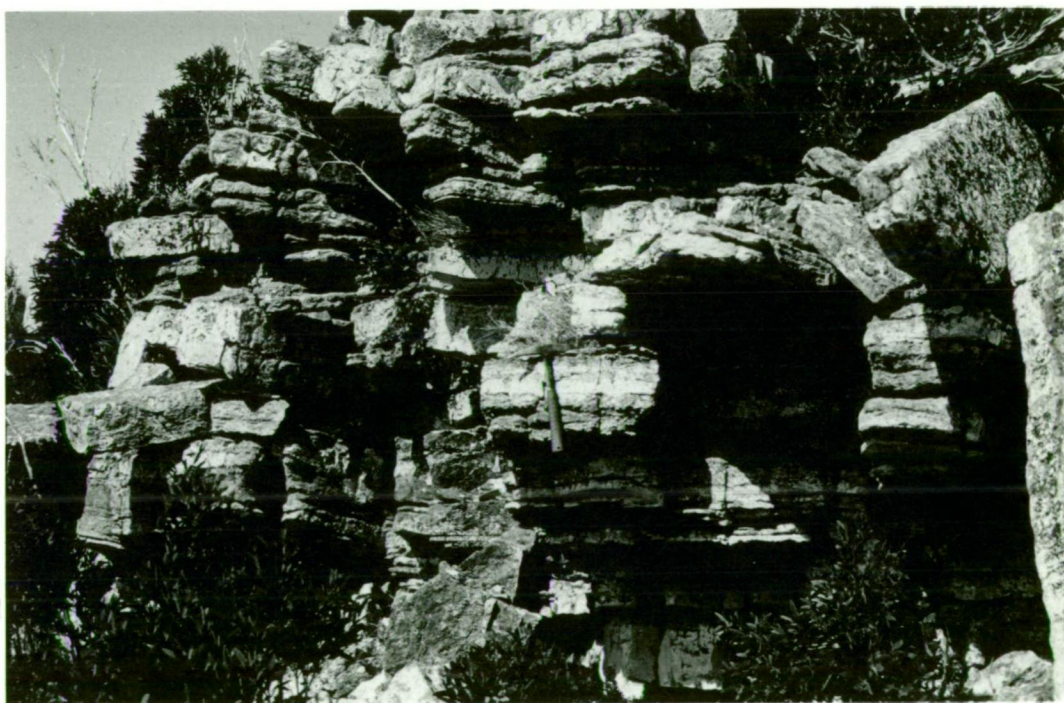
c. Tim Shea Sandstone (see Appendix A for definition)

This formation rests unconformably on the Precambrian rocks of the Jubilee Block around the eastern flank of the Florentine Synclinerium (Fig. 2). It consists mainly of quartzose sandstone and conglomeratic sandstone with lesser pebble conglomerates, and appears to be laterally equivalent to the Reeds Conglomerate. It shows considerable lateral variations in thickness, the maximum thickness being about 1000 feet (300 m) at Tim Shea and near Mt. Mueller (Figs. 2, 6). The thickening in the latter area appears to be related to the small Cambrian trough at Mt. Mueller.

The base of the formation at Tim Shea is a very irregular unconformity on Precambrian dolomite, with at least one channel-like



A. Red facies of Tim Shea Sandstone, near crest of Tim Shea. Conglomeratic red sandstone with abundant trough cross-bedding.



B. Grey facies of Tim Shea Sandstone, north of Mt. Mueller. Flat-bedded grey sandstone showing bioturbation and abundant worm burrows.

feature up to 200 feet deep. The channel is filled with dolomitic breccia, and this is overlain by poorly exposed dolomitic sandstone followed by about 40 feet of red siltstone containing abundant small worm tubes. These pass upwards into pink sandstones.

Characteristically, the main part of the formation consists of alternating zones, up to several hundred feet thick, of red cross-bedded sandstones (Plate 3A) and grey, flat-bedded worm-burrowed sandstones (Plate 3B). Red sandstones predominate at Tim Shea, while at Mt. Mueller the sequence is mostly grey. Poorly-preserved gastropods occur in grey sandstones near the base of the sequence in the latter area, and the grey horizons in all areas appear to be marine. The red sandstones, on the other hand, do not appear to contain organic traces, and are probably non-marine. A flat coastal plain seaward of the large alluvial fans of the Reeds Conglomerate seems the most likely environment.

CHAPTER 2

SEDIMENTOLOGY OF THE SINGING CREEK SILTSTONE

A. GENERAL

The Singing Creek Siltstone forms the lower part of the Upper Cambrian Denison Group, and is about 2600 feet (780 m) thick in the type area on the Denison Range (Fig. 5). It is a flysch-type deposit consisting mainly of micaceous siltstones and very quartzose greywacke sandstones. A composite section through the formation is shown in Fig. 7. The basal conglomerates are poorly exposed and are not included in this analysis. Best exposures of the main part of the formation are around Flagstone Knoll, a bare, steep-sided feature on the lower western flank of the range, and most of the following account is based on observations in this area. Outcrops elsewhere are very limited because of the thick scrub and extensive talus deposits, but the general features of the formation appear to be similar in all areas.

For descriptive purposes it is possible to classify the primary rock types into four overlapping groups which are used as a basis for sedimentological analysis, viz. (A) laminated siltstones and fine sandstones, (B) thin-bedded sandstones (usually graded), (C) thick-bedded sandstones (commonly non-graded), and (D) conglomerates. The field evidence suggests that all these rock types were deposited by the same kind of current. Slump sheets, in which some or all of these primary rock types are mixed together, form another important component and are discussed under Group E.

B. ARRANGEMENT OF ROCK TYPES

The three-dimensional relationships between the various rock types within the formation are difficult to study because of the nature of the outcrops and the terrain. The outcrops are mostly in the form of small patches on steep slopes, and are separated by large areas without exposure. The complex stratigraphy, lack of marker horizons, and apparent lateral impersistence of many of the units makes extrapolation between outcrops extremely uncertain. Sections

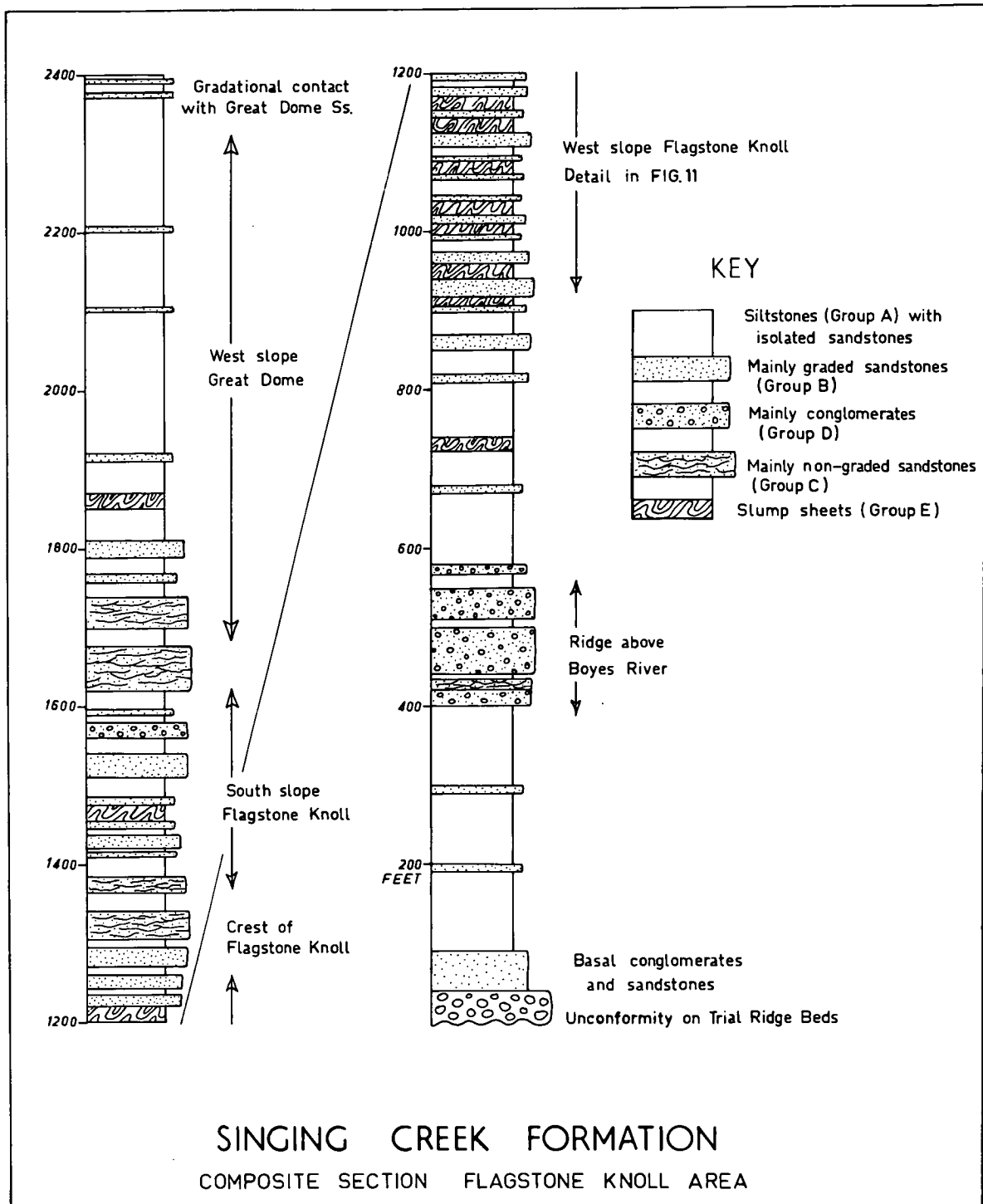


FIGURE 7

only a few hundred feet apart cannot be matched with confidence, and it is apparent that many beds, particularly the sandstones and conglomerates, are either very variable or impersistent. This is also indicated by the lateral variations shown by many coarse-grained units across single outcrops. Detailed surveying of the best exposed areas, accompanied by tracing and matching of as many units as possible, seems to be the best solution, but surveying is difficult because of the steep slopes and the general access problems.

In the type section at Flagstone Knoll (Fig. 7) the sandstones and conglomerates occur sporadically throughout the formation but the majority are contained within two major intervals. The lower interval, which is about 150 feet (45 m) thick, forms the crest of a small spur above the Boyes River and consists predominantly of conglomeratic beds. The upper interval is about 1000 feet (300 m) thick and underlies about 600 feet of siltstones at the top of the formation. There is a general concentration of sandstone beds in the upper half of the formation for several miles to the north and south of the type area, forming a series of small knolls (Fig. 5). In the Staircase Rocks area, just to the north, however, the top of this interval is only 200 feet below the top of the formation.

Within the upper interval there are many single sandstone beds isolated within siltstone sequences, but there is also a pronounced vertical alternation of zones consisting almost wholly of sandstone/conglomerate with zones consisting mainly of siltstone. This alternating pattern of coarse- and fine-grained zones is recognizable in most areas, although the siltstone units are generally poorly exposed and tend to form gullies or low areas. The coarse-grained zones range in thickness from about 5 feet (1.5 m) up to about 80 feet (24 m), and the intervening siltstone horizons are of the same order of thickness (Fig. 7). Many of the siltstone zones have been affected by slumping, and in some places, as on the western slope of Flagstone Knoll (Fig. 11), practically all the fine-grained horizons have been incorporated into slump sheets.

Many of the coarse-grained zones form small ridges of outcrop projecting through the cover of talus and vegetation on the small

knolls and down some of the steep gullies along the range. Their lateral persistence is difficult to demonstrate, however, and at least some do not persist in recognizable form for more than a few hundred yards. Many others cannot be recognized with certainty in sections half a mile away along strike, and the general impression is that the coarse-grained zones are actually lenses sitting in a "matrix" of siltstones. For example, at Staircase Rocks the upper 600 feet of the formation includes at least three major coarse-grained zones, while half a mile to the south at Flagstone Knoll this part of the formation consists wholly of siltstones. Similarly, a number of the coarse units on the south flank of Flagstone Knoll and in the Singing Creek area cannot be traced down the north flank. The edges of the units are not exposed, however, and it is not known if they are abrupt or interfingering with the siltstones.

C. GROUP A - SILTSTONES

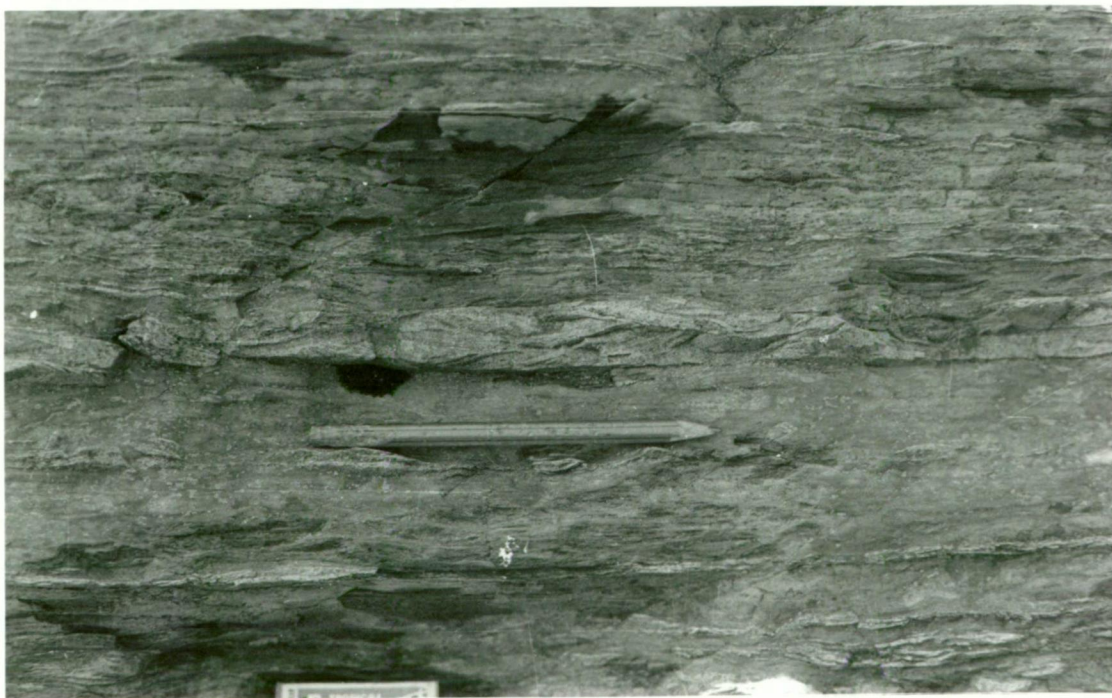
Rocks of this group constitute roughly half of the formation. They are generally poorly exposed except in isolated patches on the lower western slopes of Great Dome and around Flagstone Knoll. Outcrops are mostly buff-coloured due to limonite-staining, but fresh specimens are medium to dark grey. Most of the limonite is apparently derived from weathering of authigenic iron-bearing carbonate. Lamination is prominent in nearly every outcrop. Fossils occur within the siltstones in many places, and are abundant on some horizons. The most common types are small inarticulate brachiopods, agnostid and ptychopariid trilobites, and small orthid brachiopods. Sand-filled burrows up to about 5 mm in diameter are common, and rare arthropod (trilobite ?) tracks have also been seen. Mica is abundant on all stratification surfaces, and is the dominant constituent of the finer-grained layers.

1. Lamination and Graded Bedding

The prominent lamination is formed by alternations of light-coloured quartzose bands with dark micaceous silty bands on a scale of less than one mm to several cms (Plates 4, 5, 6). In the



A. Graded sandy layers within Group A siltstones. Note sharp bases, gradational tops, division of deformed lamination. Scale in inches.



B. Cross-laminated and ripple-marked layers within Group A siltstones. Note semi-isolated ripple lenses beneath pencil; small normal fault cutting sandstone layer.

fossiliferous horizons the lamination is commonly disrupted or almost completely destroyed by burrowing, giving a mottled structure. The quartzose bands consist mainly of very fine sand, and are usually thinner than the silty ones. Thicker sandy layers are also common, however, (Plate 4A) and form a gradational series with the thin graded sandstones of Group B (e.g. Plate 7A). Graded bedding is very common, and occurs on all scales. Most of the thin quartzose laminae have sharp bases and gradational tops into micaceous siltstone (e.g. Plate 5A, B), but there are also many which are sharp-topped. The thicker quartzose laminae, such as shown in Plate 4A, are also graded, with gradational tops passing up into laminated siltstone in which the sub-laminae are also graded. In such cases it is usually impossible to decide how much of the overlying siltstone belongs to the same depositional unit as the quartzose sandy layer, and the concept of the sedimentation unit is difficult to apply to these rocks in general.

Most of the thicker quartzose layers show internal lamination, consisting of flat lamination (Plate 4A) or cross-lamination (Plates 4B, 6A) or both. Changes in grainsize and/or mica content are generally apparent between adjacent sub-laminae, and the sub-laminae themselves may be graded. Such sub-laminae become difficult to distinguish from primary units as the silt content increases (e.g. Plate 5A).

Some of the cross-laminated units have rippled tops and a flat base (Plate 6A), while some consist of a series of very small ripple lenses connected only by a line of quartz grains or actually isolated (Plate 4b). Although the tops of the rippled units are commonly sharp, graded bedding is suggested in some cases by the presence of silty foresets in the tops of the ripples (Plate 6A). An interval of flat lamination occurs beneath the rippled tops of some of the thicker layers. The ripples are usually only slightly asymmetrical, but the cross-lamination always indicates unidirectional currents. Wavelengths range from about 30-70 mm and amplitudes from 3-10 mm. Cross-laminated fine sandstones become more common towards the top of the formation at Great Dome, and there are rare units up to 12 inches (30 cm) thick showing cross-lamination throughout.

2. Deformational Structures

Some of the thicker graded units include a division of deformed lamination, usually associated with cross-lamination, which resembles the "convolute lamination" of turbidite literature. An example is shown in Plate 4A in which the contorted division consists of laminated silty fine sand and is separated along a decollement surface from the underlying lamina.

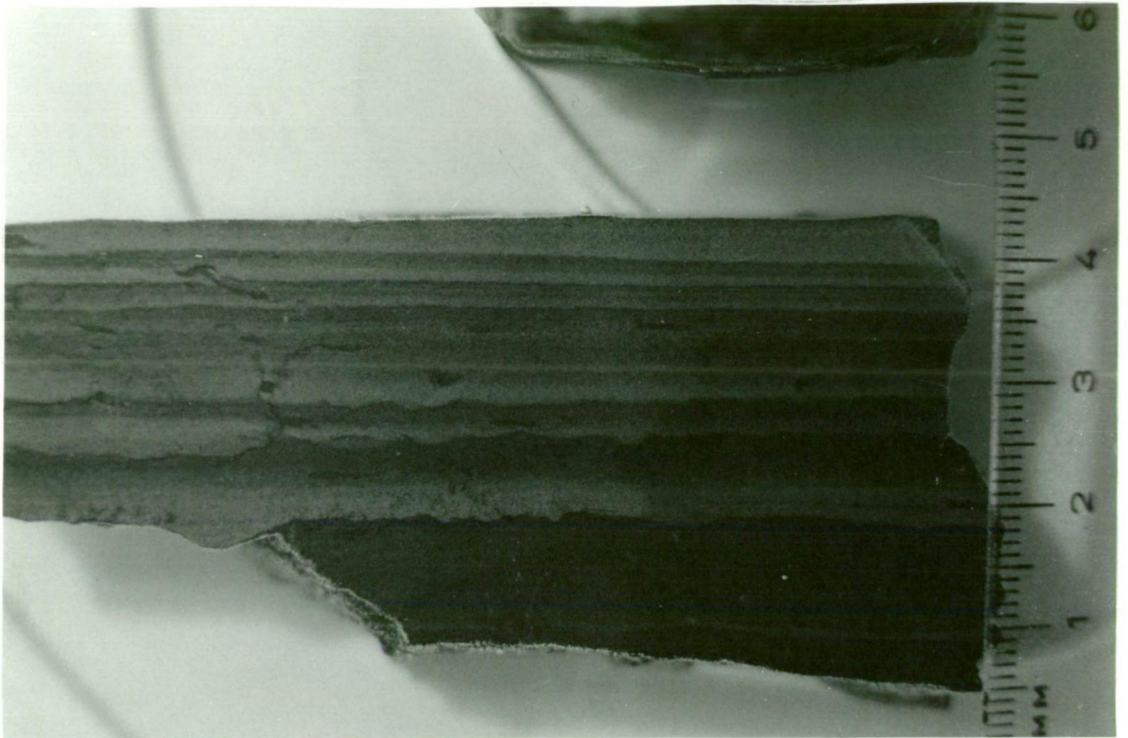
Small-scale normal faults occur in many exposures of the siltstones. The faults are generally inclined at 30° - 40° to the bedding, and dip either north or south. The amount of displacement ranges up to about 10 mm. An example is shown in Plate 4B where the dip of the fault changes to almost vertical as it cuts through a thin sandstone layer. Many of the faults die out within the limits of the outcrop, suggesting they are probably compactional features.

3. Sole Marks

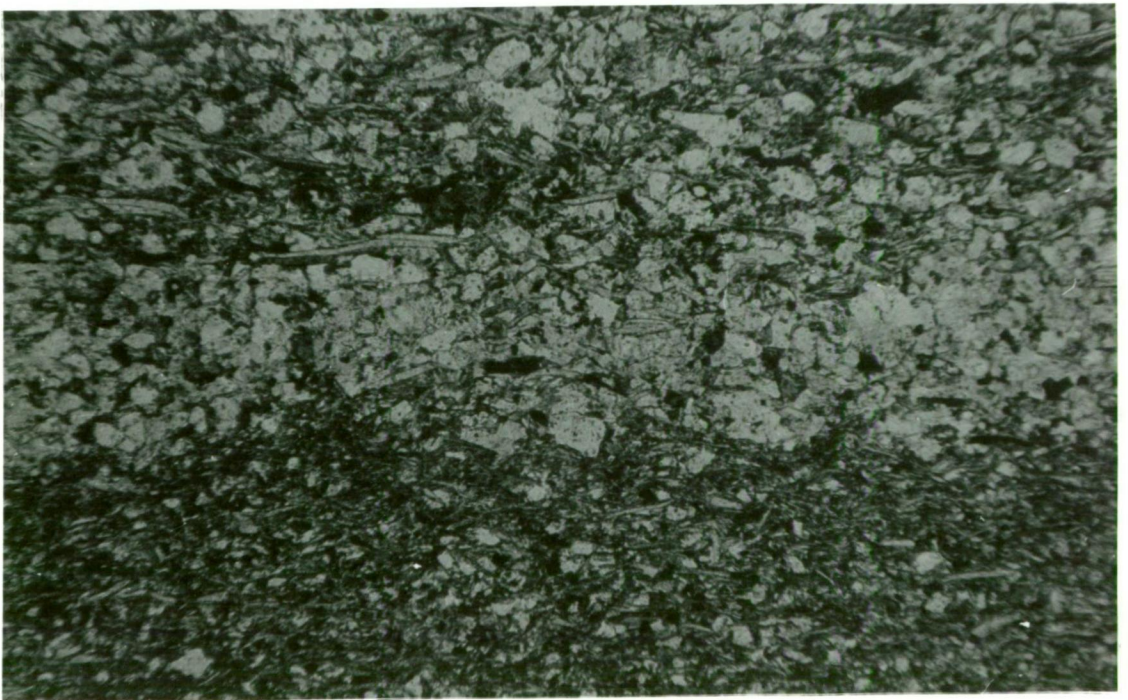
Many of the sandy layers have small-scale load marks and/or scour structures on their soles, e.g. Plate 4A (upper layer) and Plate 5A, although flat bases are more common. Some of the thicker units show well-developed longitudinal ridge and furrow structure as described for the Group B sandstones. The silt immediately beneath some of the sandy laminae shows a disturbed structure due to slight scouring, and fragments of micaceous material detached from the silt occur within some of the sandy laminae (Plate 5A). Such contacts suggest that a bottom-hugging current was involved in the deposition of the sand, and are not indicative of quiet settling from suspension.

4. Petrology

About 20 thin sections of siltstones from various parts of the formation have been studied (see Appendix B for numbers). They show a surprising uniformity of composition, and the variations seen are mostly due to variations in the relative proportions of the constituents. This same uniformity of composition is also shown by the sandstones and conglomerates, and it is apparent that the whole formation was derived from the same source area. Only three major



A. Graded quartzose laminae in Group A siltstones. Note small scour features, fragments of dark silt. Scale in centimetres.



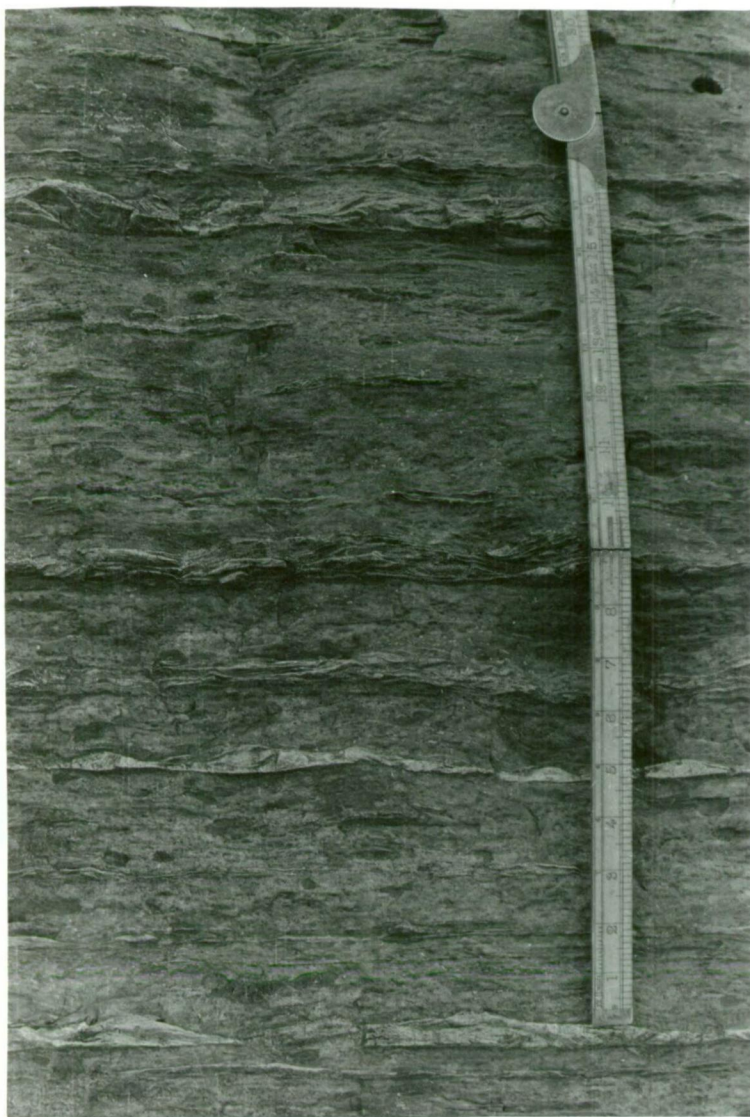
B. Photomicrograph of contact between sandy and silty laminae. Note sharp irregular base and gradational top of sandy laminae; abundant muscovite flakes in siltstone. Height of field is 1 mm.

components are present, viz. quartz grains (including undulatory and polycrystalline grains), micaceous minerals (muscovite, sericite, chlorite, biotite), and carbonate. Pyrite is a common minor constituent, while accessory heavy minerals, particularly tourmaline and zircon, are common in the sandy laminae. All variations occur from very micaceous siltstone with only scattered quartz grains to quartzose fine sandstone with a small amount of mica, and such variation commonly occurs within a single lamina.

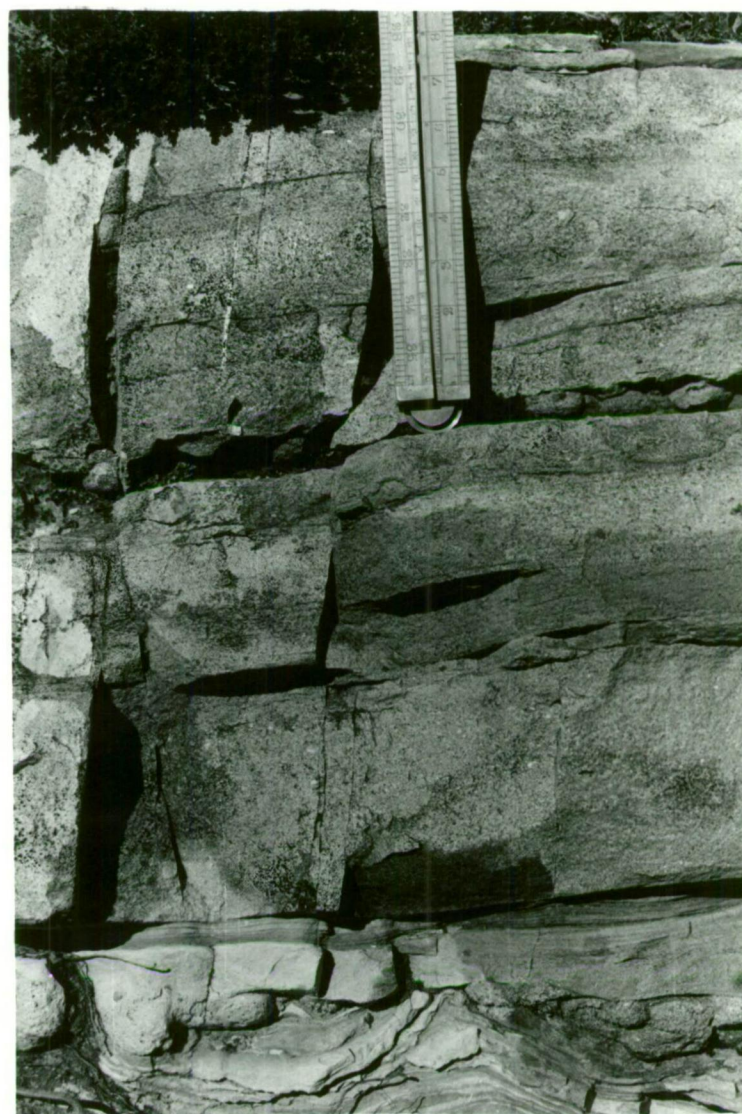
Quartz grains are common in all sections, the minimum proportion being about 10% in the very mica-rich laminae (Plate 5b). Most grains are in the coarse silt to very fine sand grade, and the maximum grain size rarely exceeds medium sand. Most of the grains are angular to sub-angular. Many of the grains (40-50%) show undulose extinction, and at least 10% are visibly polycrystalline, i.e. quartzite rock fragments. For the most part the quartz grains are separated by mica flakes, but in the sandy laminae the grains may be in direct contact with the interstices filled with authigenic silica (Plate 5B). Irregular grain boundaries due to replacement by carbonate or to interaction with sericitic material are common.

Micaceous minerals are the dominant constituents of the siltstones. Muscovite and sericite are the predominant types, making up about 70% of the micaceous component, with chlorite and biotite forming the other 30%. The muscovite flakes are up to 0.5 mm long and mostly show a strong alignment parallel to bedding (Plate 5B). Many of the flakes are bent between adjacent quartz grains. The sericite forms shredded aggregates and minute flakes and appears to be largely detrital, although a component derived from recrystallization of clay minerals is also probably present. Chlorite occurs as detrital flakes, as an alteration product of biotite, and as a fine intergrowth with some of the sericite. The dark grey colour of the siltstone is mainly due to the chlorite and biotite component.

The carbonate appears to be predominantly authigenic, and for the most part occurs as sparry crystalline blebs, up to 0.2 mm long, replacing the other constituents. Detrital grains of similar-looking carbonate occur in a few of the sandy laminae. The percentage



A. Cross-laminated and ripple-marked laminae within Group A siltstones. Scale in inches.



B. Composite sandstone sequence typical of Group B, with numerous stratification surfaces. Basal bed truncates slump sheet.

of carbonate varies greatly, from less than 5% to about 50% in a few samples, but the average is about 10%. In some samples the carbonate shows a preference for the sandy laminae, but in other cases is equally abundant in the micaceous silty layers. The carbonate is very susceptible to weathering, and weathered samples show only limonite-coated cavities from which the carbonate has been dissolved. Limonite also occurs along the cleavage traces and edges of many fresh grains, indicating that the carbonate is an iron-rich variety. Stain tests indicate ankerite and/or siderite.

Pyrite is present in nearly all samples as very minute spherulites and crystals disseminated through the micaceous material. For the most part the grains are less than 0.015 mm diameter, but aggregates may be up to 0.1 mm long. Very rare crystals reach 0.5 mm and are just visible to the naked eye. In weathered specimens the pyrite is pseudomorphed by red translucent haematite.

5. Deposition of the Siltstones

Little is known in detail about the deposition of fine-grained sediments, particularly those in flysch sequences. Most authors suggest two possibilities, viz. quiet settling from suspension, or deposition by weak bottom currents of some kind (including turbidity currents). In the present case the abundance of cross-lamination and flat lamination, and the presence of scour structures beneath some layers, suggest that bottom currents must have been involved, at least in the formation of the quartzose laminae. The fact that most of the coarser laminae grade into micaceous siltstone suggests that much or all of the mica was transported and deposited by these same bottom currents. The sharp bases and graded bedding shown by most of the laminae indicate spasmodic waning currents or pulses of some kind.

There appears to be a gradational series from the thin graded layers such as shown in Plate 4A to the typical sandstones of Group B (e.g. Plate 7A). As well as this there are many examples of thick sandstone layers which grade upwards into typical laminated siltstone, and a few examples where sandstones pass laterally into

siltstones (Fig. 8). It would seem, therefore, that the deposition of at least some of the siltstone was accomplished by the same currents which deposited the sandstone layers. As discussed later, the sandstones were apparently deposited by bottom-hugging density underflows (i.e. turbidity currents), and much of the siltstone also appears to be of this origin.

The reason for the micro-grading of the very thin laminae and sub-laminae is not immediately apparent, but it seems hardly likely that each of these represents a separate major density underflow. The division of laminated siltstone which forms the upper part of many graded sandstone beds may include a large number of these micro-graded laminae, indicating that each major current was capable of producing a number of such laminae. Walker (1965) has discussed this problem and suggested that graded laminae in turbidites could develop "by a process of intermittent supply of mixed sediment to the top of the laminar boundary layer". The laminar boundary layer is a thin fluid film which moves with laminar flow between the smooth bed surface and the over-riding turbulent fluid (turbidity current) at low velocities. The grading was thought to develop because of differential settling of various grain sizes through this layer. While this may be so, there must be some other reason for the intermittent supply of sediment to form the laminae, since a steady supply would not result in grading. Eddies or vortices within the turbulent flow, successively impinging on the bottom as the overall velocity of the flow decreases, seems a better general explanation.

The probable importance of turbidity currents as primary agents for transporting and depositing the fine-grained sediment in flysch sequences has been emphasized by several authors (e.g. Pässega, 1954; Radomski, 1960; Ksiazkiewicz, 1961). Dzulynski and Walton (1965, p.11) suggested that secondary currents produced from the action of major currents on the muddy seafloor, and forming overflows, interflows or underflows, would also be important. Crowell et al. (1966, p.38) postulated that "weak evanescent filaments, reaching out from the main spatulate mass of the underflow" deposited much of the mud between the sandstone beds, and this agrees with the present

author's conclusions. As discussed later, the underflows were apparently complex, laterally-variable features, some of which deposited thick sand or gravel layers in some places and a thinner blanket of laminated silts over surrounding areas.

The possibility of indigenous ocean-bottom currents must also be considered. Present knowledge suggests, however, that while these currents may be capable of reworking and rippling the bottom sediment, they probably do not transport large amounts of sediment into ocean basins (e.g. Hsu, 1964). The nature of the lamination produced by such currents is not known, and it is doubtful if their effects could be differentiated from those of density currents in fine-grained sediments. The apparent increase in the abundance of cross-lamination towards the top of the formation suggests that ocean currents may have been more influential during this period, possibly reflecting the shallowing of the basin as the shallow marine facies of the Great Dome Sandstone advanced. Although very poorly exposed, it is possible that the laminated rocks towards the top of the formation represent pro-delta deposits.

D. GROUP B - THIN-BEDDED SANDSTONES

1. General Features

Most of the sandstones of the Singing Creek Formation occur as beds less than 12 inches (30 cm) thick, and are included under Group B. It is emphasized, however, that there are all gradations from these sandstones to both the thick-bedded sandstones of Group C and the conglomerates of Group D, as well as to sandy laminae of the Group A siltstones, and there is considerable overlap between all these groups.

The sandstones are hard, light-coloured, very quartzose rocks which in outcrop might be called quartzites. This is somewhat unusual since there is abundant evidence that the beds are actually turbidites. Shale pellets and transported fossils (brachiopods and trilobites) occur in some beds. Grain size varies from pebble conglomerate to very fine sandstone or siltstone, but most beds are predominantly medium to fine sand grade. Many beds have a conglomeratic

division at the base and are obviously graded. Tops of the beds are generally sharp, and may be overlain by another sandstone bed or by laminated siltstone. In some cases there is a gradational passage from the sandstone to the siltstone, indicating that both were formed by the same depositional episode as is the case with most flysch beds. Generally, however, there is a sharp separation between the two which is accentuated by weathering (e.g. Plate 7A), and it is seldom obvious that both rock types are part of the same bed.

Weathered surfaces of the sandstone are bleached almost to white to a depth of several cm, and beneath this there is usually a brownish limonite-stained zone from which the carbonate has been dissolved. The fresh rock is medium grey in colour. Disintegration of outcrops occurs by splitting along joints and stratification surfaces, producing slab-like blocks which accumulate as hard platy scree on many slopes.

The sandstones occur as isolated beds within siltstone sequences or, more usually, as clusters or groups of beds separated by zones consisting mainly of siltstone, as discussed previously. The groups of beds range in thickness from a few feet up to several tens of feet, and in some cases form small ridges. Interbedding and intergrading with thick-bedded sandstones and conglomerates occurs in many of these coarse-grained zones.

2. Stratification

a. Bed Thickness

There is a complete range in bed thickness from the laminae and very thin beds described under Group A up to beds about six feet (2 m) thick of Group C, but the average thickness for Group B is estimated to be 5-8 inches (12-20 cm). The thick sandstone layers which are not Group C beds can usually be shown to be composite, and as many as 20 amalgamated "beds" may be present within a sandstone layer several feet thick. Beds of any thickness may occur within a few feet of section, but a tendency for a number of beds of similar thickness to occur together is apparent in many places (e.g. Plate 7A).



A. Group B sandstones. Note sharp tops and bottoms; ripple marks; tendency for beds of similar thickness to form groups.



B. Compositely-bedded sandstone sequence showing profusion of stratification surfaces. Truncated slump sheet beneath hammer.

Although some authors have noted a direct relationship between bed thickness and grainsize in flysch formations (e.g. Crowell et al., 1966), such a relationship is by no means obvious in this sequence. Much of the conglomerate occurs as thin beds or as bands or lenses within sandstone beds, while most of the thicker sandstones are not noticeably coarser than the majority of the thin beds. The relationship appears to apply only at the ends of the scale, in that the thickest beds of the sequence are conglomerates and the thinnest beds are siltstones.

Most of the isolated sandstone beds maintain a fairly constant thickness across the outcrop width (usually less than 200 feet or 60 m), but within the compositely-bedded sequences there may be many variations in bed thickness. Some sandstone layers show remarkable lateral variations in both thickness and grainsize, some examples of which are discussed later.

b. Types of Stratification Surface

The compositely-bedded sandstone horizons commonly show a profusion of sub-horizontal stratification surfaces wherein bedding surfaces are difficult to distinguish from internal lamination planes, and primary depositional units are almost impossible to define (Plates 6B, 7B). Measurement of bed thickness in such sequences becomes very subjective, and rigid rules are difficult to apply. In three-dimensional exposures the internal lamination planes can sometimes be distinguished because they are either smooth or show parting lineation, whereas the bedding surfaces are usually rough-textured due to the presence of coarse grains or fossil fragments, and may show sole marks (e.g. Plate 7B). Some of the rough-textured surfaces become smooth laterally, however, and some die out altogether. Lateral passage of conglomeratic bands into thin smooth joints, such as shown in Plate 8A, suggests they could represent lamination planes rather than bedding surfaces.

Current marks and deformational structures may be used to identify bedding surfaces in composite sequences. The current marks are seldom as well defined as they are on beds overlying siltstones or



A. Sandstone layer between two slump sheets. Note basal conglomeratic division, also pebbly layer at base of overlying slump; internal conglomeratic bands pass laterally into thin joints.



B. Same unit as above, about 100 feet further south. Most of sandstone layer has been incorporated into overlying slump sheet. Note sharp contact between conglomeratic division and laminated sandstone.

slump sheets, but include flute marks, groove marks, obstacle scours (particularly around large quartz grains), and rarely longitudinal ridges and furrows. Deformational structures include load pockets, often with coarse-grained fillings (e.g. Plate 10B), and cusp structures consisting of pointed "anticlines" separated by broad basin-shaped "synclines". Structures of the latter type appear to have been formed by a combination of loading and extrusion of water (through the anticlines) from the underlying saturated sand.

In many such sequences it is difficult to decide if the bedding surfaces (as opposed to lamination planes) which are visible represent the boundaries of complete depositional units or of sub-units formed by pulses within a single current. This problem has also been raised by Hubert et al. (1966), who described a single "bed" in which several "intra-bed surfaces" show groove moulds (and the shale fragments which cut them) similar to those on the base of the bed. As stated by these authors (ibid, p.241) such features "could be the result of a single depositional influx with pulsating sedimentation, or multiple independent depositional events". Internal lamination planes, on the other hand, were considered to mark only minor fluctuations in the depositing current. There is thus a suggestion of a hierarchy of surfaces, from the primary bed surface through "intra-bed surfaces" to internal lamination planes, reflecting a decreasing scale of depositional episodes. This could perhaps be used as a basis for classifying stratification surfaces, but the distinction between primary bed surfaces and intra-bed surfaces may be difficult or impossible in composite sandstone sequences.

An example of what appears to be a single "bed" showing a scoured intra-bed surface is shown in Plate 9A. Here a sandstone layer about 12 inches (30 cm) thick separates two slump sheets. The layer has flute marks and shallow channels on its base where it truncates the slump sheet, and another surface about 3 inches above the base also shows flute marks. There is a suggestion of a third erosional surface just above this again.

Surfaces which appear to be intermediate between intra-bed surfaces and internal lamination planes also occur within some beds.



A. Sandstone layer between two incoherent slump sheets. Note scoured intra-bed surface above base. Rule is 9 inches long.



B. Sandstone layer with truncated flute marks. Some of the flutes have coarse-grained filling. Layer truncates incoherent slump sheet.

For example, the bed shown in Plate 9B has flute marks of various sizes along its base, where it truncates a slump sheet. The flutes are isolated and are separated by a flat stratification surface which continues across the top of the flutes and appears to truncate them. Some of the flutes contain sediment which is much coarser than that above the truncating surface, but in other cases there is no marked change in grainsize across the surface. This suggests that all the sediment was deposited during the same overall episode, but that the flutes were cut and filled (or partly filled) by an early phase which was followed by a second phase which truncated some of the flutes and deposited the bulk of the bed.

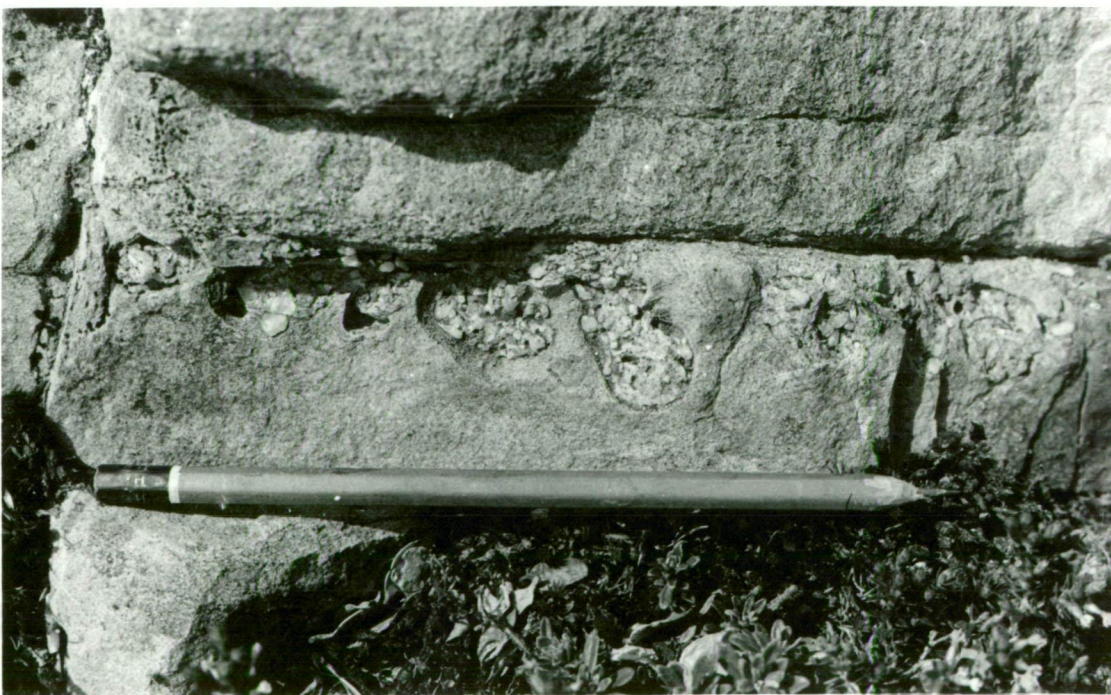
Similar examples of beds which have flute marks and other sole structures filled with coarse material and apparently truncated by an intra-bed surface of some kind are fairly common in the Group B sandstones. Two other examples are shown in Plates 6B (near base) and 10A. The flute marks shown in Plate 14B are also of this type. In nearly every case the sediment within the flutes is both unsorted and unlaminated whereas that above them is distinctly laminated and better sorted. This suggests that the second phase of the current was somewhat different in nature from the early phase. Beds in which the unsorted conglomeratic material is spread along the base as a more or less uniform layer and overlain along a distinct surface by laminated and better sorted sandstone, are also fairly common (e.g. Plate 8A, B).

3. Examples of Lateral Variations

Many beds show marked lateral variations in thickness and/or grainsize. Thickness changes are commonly due to the presence of deep sole marks along the base of an otherwise thin bed, e.g. Plate 11A shows a sandstone layer which is 12 inches (30 cm) thick where it fills a channel structure, but which thins to almost zero within about 3 feet laterally. Many similar examples in which sandstone layers double or treble in thickness due to the filling of deep sole marks can be found, and in many such cases the thickening is accompanied by an increase in grainsize due to the trapping of coarse material in the depressions.



A. Sandstone layer with truncated flute mark at base. Overlies semi-coherent slump sheet with "stretched" bedding. Overlain by incoherent slump sheet with large sandstone balls.

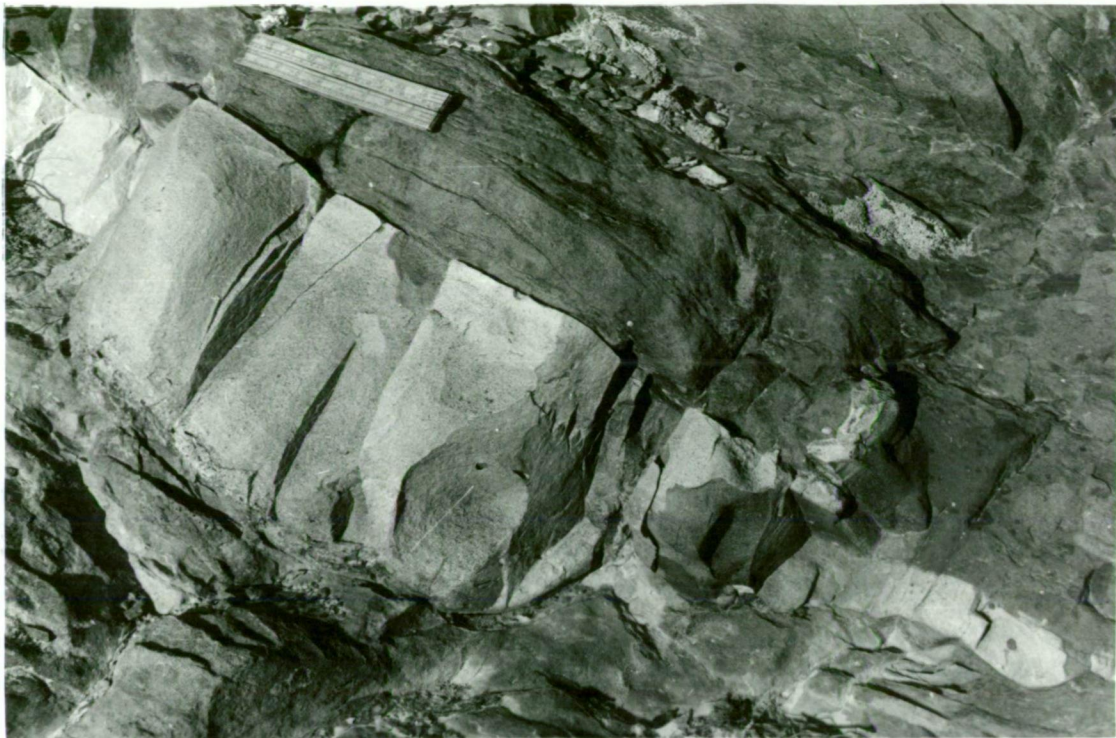


B. Load pockets filled with coarse grit developed at contact between two sandstone beds.

There are other examples, however, in which the variations are apparently not due to the development of sole marks. Careful tracing of beds, or very good exposures, are usually necessary to show up such changes. In the example shown in Fig. 8 and in Plates 11B; 12A, B, a sequence of several beds can be traced continuously along strike for 25 feet (8 m). The outcrop surface trends north-south, which is almost perpendicular to the current direction, and hence the variations are likely to be lateral rather than longitudinal with respect to the original current. The basal bed truncates a slump sheet with prominent flow cleavage (No. 26 in Fig. 11). Fig. 8 shows this sequence drawn to scale, based on sections measured every few feet and horizontal tracing of beds.

The basal bed (Bed A) consists mainly of laminated sandy siltstone, with a thin discontinuous lamina of fine sandstone at the base in many places (Plate 11B). This lamina is ripple marked and cross-laminated and is similar to the fine sandstone bands in the Group A siltstones elsewhere. It is thicker and coarser-grained at the southern end, where some small scour marks contain grit and small pebbles (Plate 12B). In the middle part of the exposure, however, the basal unit is formed by a flat mound or lens of sandy conglomerate (Plate 12A) containing pebbles up to 2 inches (5 cm) long. The conglomeratic lens is about 7 feet (2 m) wide and 3 inches (7 cm) thick. At the edges of the lens the conglomerate passes into rippled sandstone which is continuous with the sandstone lamina at the base of the bed elsewhere (Plate 12A). The top of the lens is truncated by the overlying bed. A single isolated quartz pebble, 3 inches (7.5 cm) long, rests on top of the slump sheet a few feet north of the conglomerate lens (Fig. 8), and is overlapped by the siltstone laminae. The main gravel- and sand-bearing part of the current which deposited this bed appears to have passed over this area, leaving only the lens of gravel and an isolated pebble, the bulk of the bed being deposited by the silt-bearing tail of the current.

The second unit (Beds B1, B2) comprises two sandstone layers separated by a prominent stratification surface (intra-bed surface ?) towards the top. At the southern end of the exposure this



A. Sandstone layer showing thickness variation due to filling of large channel structure. Separates two incoherent slump sheets. Note lamination extending almost to base of channel.



B. North end of the outcrop shown in Fig. 8. Note silty Bed A truncating slump sheet with flow cleavage; increase in pebble content of Bed B1 to south; thin sandy lamina at base of Bed C.

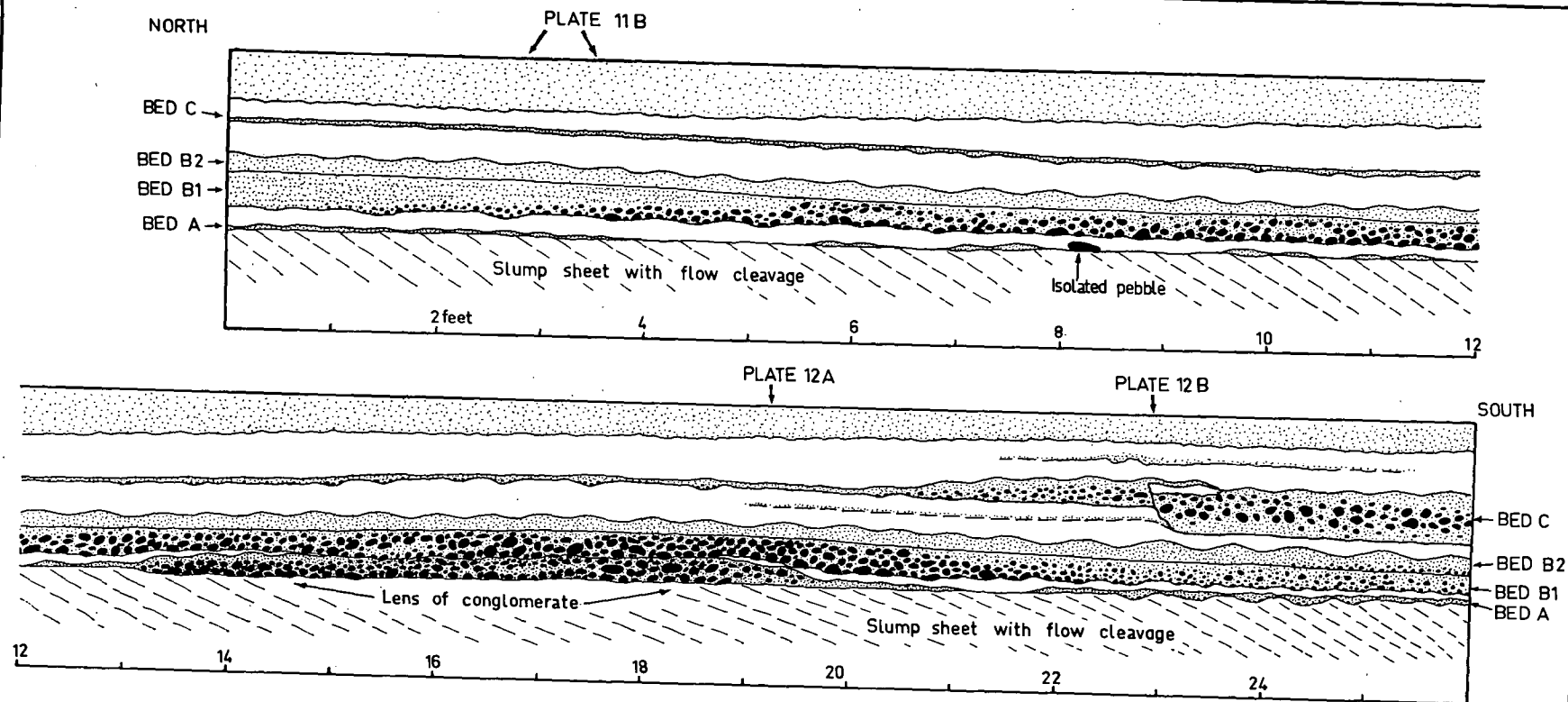


FIG. 8 SCALE SECTION SHOWING LATERAL VARIATIONS IN THREE SANDSTONE BEDS



A. Conglomerate lens of Bed A passes laterally into thin sand lamina; Bed B1 is markedly conglomeratic; Bed C begins to thicken to the south. Part of the outcrop shown in Fig. 8.



B. South end of outcrop shown in Fig. 8. Rapid thickening of Bed C due to erosion of underlying silty sandstone, with subsidence on micro-fault. Bed B1 mainly sand; Bed A has grit-filled pockets.

surface is overlain by a thin layer of conglomerate (Fig. 8). The lower "bed" is about 4 inches (10 cm) thick and is markedly conglomeratic across most of the exposure, containing quartz pebbles up to 2.5 inches (6 cm) in diameter. The amount of conglomeratic material decreases to both north and south, however, and there is virtually none at the northern end of the outcrop (Plate 11B). A fairly prominent parallel lamination is present within the sandstone part of this bed, and extends to the base of the bed where there is no conglomerate. The upper part of this couplet (Bed B2) consists of ripple-marked and cross-laminated sandstone passing upwards gradationally (in places abruptly) into laminated sandy siltstone. It is possible that the conglomeratic part of this bed originally filled a shallow depression which is no longer apparent because of compaction over the underlying conglomerate lens.

Bed C is very prominent at the southern end of the section, where the sandy part is about 7 inches (18 cm) thick and is markedly conglomeratic (Plate 12B). Further north, however, the sandstone part of this bed appears only as a thin lamina (2-6 mm) of coarse sand which grades up into laminated sandy siltstone (Plate 11B). In the middle part of the section the sandy layer consists of a series of small grit-filled hollows overlain and connected by a thin lamina of fine sand (Plate 12A). The thickness of the sandstone increases to 7.5 cm (3 inches) over a few feet, at which point there is an abrupt thickening to 12.5 cm (5 inches), as shown in Plate 12B. This is accomplished by the complete erosion of an underlying graded silty sandstone layer for a distance of at least 3 feet laterally. The erosion surface is very steep and consists of two sub-vertical steps. A coarsening of grain-size of Bed C occurs to the right of this step.

Syn-depositional subsidence of Bed C

The upper surface of the sandstone part of Bed C is displaced sharply downwards at this erosional step by about 2 cm along a steep micro-fault (Plate 12B; Fig. 8), and the hollow so formed on the down-thrown side is filled with siltstone. This in turn is covered by the laminated fine sandstone which elsewhere rests directly on the coarse sandy part. The micro-fault does not persist through this

fine sand lamina or into the overlying siltstone. The explanation for this seems to be that the coarse sandy layer subsided immediately after being deposited and prior to the deposition of the fine sand lamina, forming a pocket which for some reason trapped silt from the current rather than sand.

The subsidence appears to have been largely due to compaction of the underlying silt, particularly in the hollow between two ripple crests beneath the erosional step. The greater weight of the thickened part of Bed C, and the erosion of the supporting silty sandstone layer, probably induced the compaction. The down-warping of the truncated end of this silty sandstone layer corresponds to the fault movement in Bed C. The thickening of the gravelly part of Bed C within this eroded trough or channel suggests that deposition from the initial part of the current was strongly influenced by bottom topography so that it occurred mainly in the depression.

Although such well-exposed examples are rare, there must be many cases of similar variations within sandstone beds. This becomes apparent when attempts are made to correlate between adjacent outcrops, including those only a few tens of feet apart. While it is usually possible to correlate one or two of the beds, the identification of others is usually difficult or impossible because of lateral variations. This is particularly so in those beds which contain conglomeratic material, since this tends to be distributed irregularly along the bed. Those beds overlying slump sheets commonly show pronounced variations as do those within the compositely-bedded sequences.

4. Sole Marks

The sandstones show a variety of sole marks, although the majority of beds, and particularly those resting on other sandstones, have flat soles. The nature of the pre-existing surface, i.e. whether it was an irregular slump sheet or a relatively smooth sand bed, appears to have been important in controlling the development of sole marks. Current scour marks and load structures are the most common, and it is often difficult to distinguish between the two when the structures are seen only in section. Most of the scour marks show

some deformation due to sinking into the underlying mud, and in some cases are very irregular in shape (Plate 13A). Careful examination of such structures usually shows that some of the underlying laminae have been truncated, demonstrating that the structures are primarily erosional.

a. Current Marks

The nomenclature proposed by Dzulynski and Walton (1965) is used in the following discussion. The correlation noted by these authors (p.141) between the grainsize of beds and the types of current marks developed is also apparent to some extent in the Group B sandstones, viz. coarse-grained beds show mainly channels and isolated flute marks, intermediate beds show mainly patterned flute marks, and fine-grained beds show mainly longitudinal ridges and furrows, small flute marks and tool marks.

(i) Longitudinal ridges and furrows

Many thin beds, including some less than one cm thick, show this structure. The furrows (actually ridges as they appear on the soles) range in width from 1-10 mm, but are generally fairly constant in size on any particular bed. Several examples of a similar structure in which the pattern of furrows has been modified by pre-existing ripple marks have also been found (Plate 14A). In this structure, termed "modified ripple mark" by Craig and Walton (1962; see also Dzulynski and Walton, 1965, p.61), the elongate furrows are replaced by a cellular pattern along well-marked zones which are perpendicular to the current direction. These zones correspond to the crests of pre-existing transverse ripples. The cellular pattern persists into the troughs of the ripples but reverts to a longitudinal pattern up the stoss side to the next crest.

Longitudinal ridges and furrows are well-known turbidite features, and have been interpreted by Dzulynski and Walton (1965) as current scours. The formation of the furrows is apparently due to stringer-like vortices within the flow: "the stringers are organized longitudinally in the flow and in each one the fluid rotates around two helical spirals which show an opposite sense of motion. Scouring



A. Load-deformed flute marks on base of sandstone layer separating two incoherent slump sheets. Structures near right-hand end apparently affected by translation of bed from left to right.



B. Flute marks filled with cross-laminated fine sand. Same bed as Plate 11A. Note truncation of flow cleavage in underlying slump sheet.

takes place within the stringer and the eroded material is piled up on the side as a longitudinal ridge" (*ibid*, p.66). The formation of the stringers appears to be a result of density-controlled stratification within the flow, and hence the structure itself may prove to be unique to turbidites. If the flow is slowed or stopped, the horizontal pattern may be replaced by a standing pattern of cell-like vortices, producing non-directional structures such as found in the troughs of the "modified ripple mark" described above.

(ii) Obstacle scours

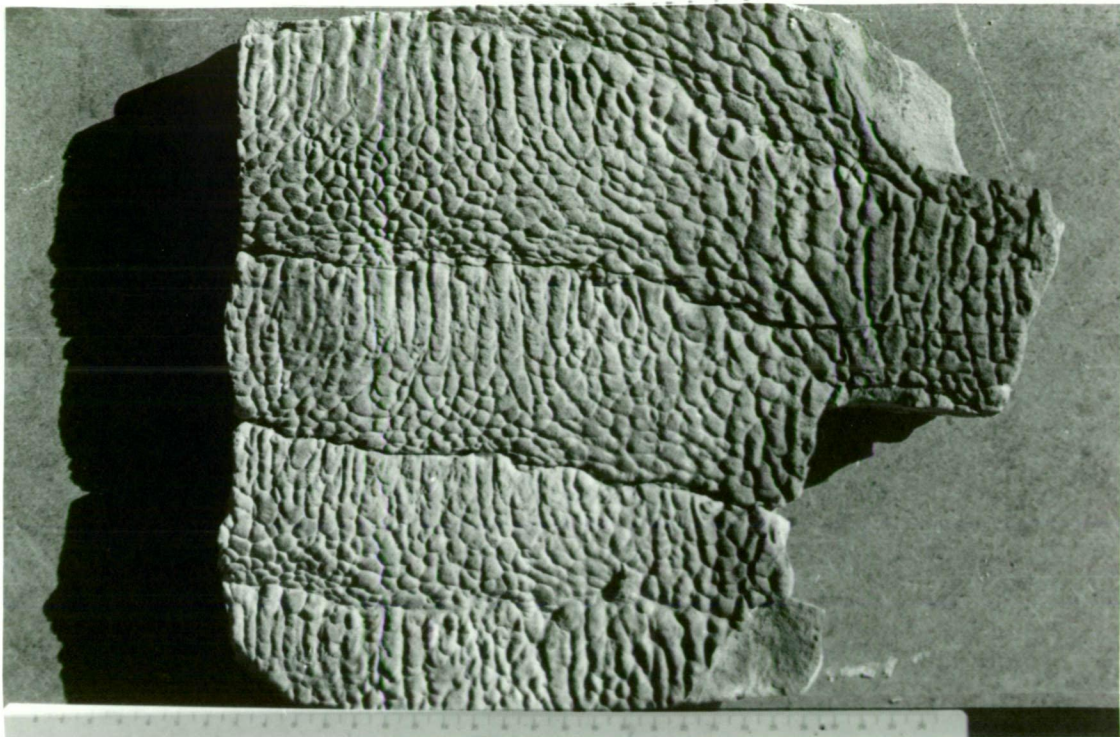
These are generally rare, and are mostly of the current crescent type in which two parallel furrows (ridges as seen on the soles) extend downcurrent from the edges of the obstacle. In some cases the obstacles consist of worm tubes 1-5 mm in diameter. Most of the tubes are perpendicular to bedding, but some are bent so that they lie on the bedding surface and a few have apparently been broken off and rolled or pushed along the bed. Large quartz grains form the obstacles on some soles, particularly those which show isolated flute marks (e.g. Plate 14B). In such cases the large grains must have been deposited by the current at some stage prior to the formation of the obstacle scours, probably during the same phase as the deposition of gravelly material in the flute marks.

(iii) Tool marks

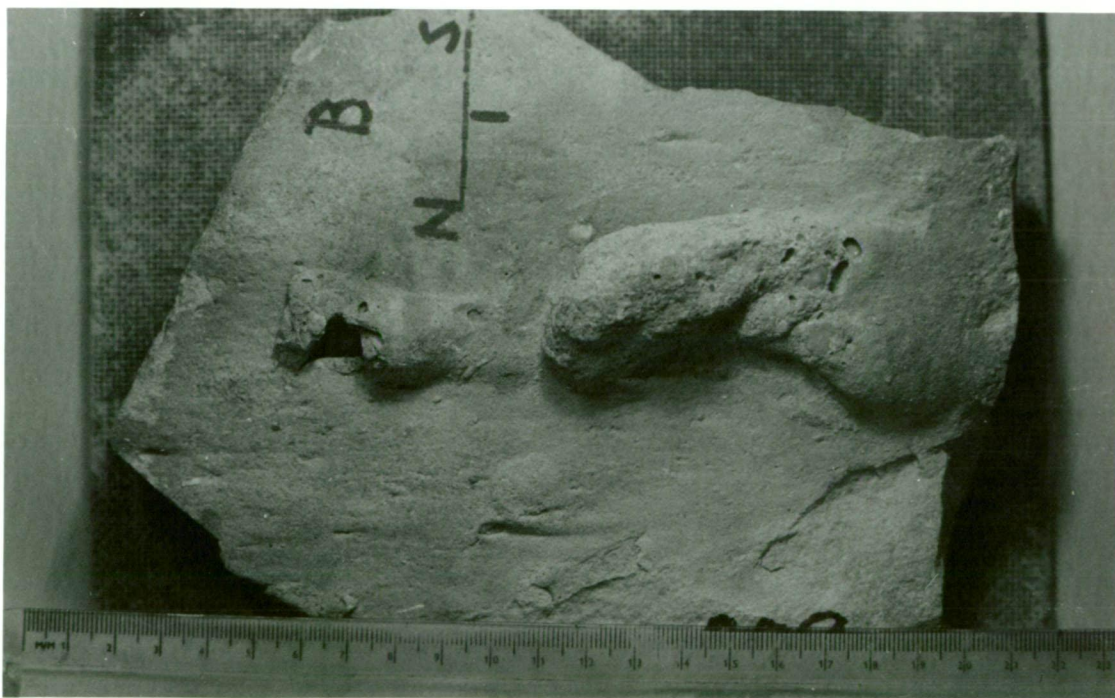
These are also rare, perhaps reflecting a general scarcity of suitable tools. Small linear impact marks are apparent on some soles resting on siltstone, and were probably formed by saltating sand grains.

(iv) Flute marks

Flute marks are common, particularly on those beds overlying slump sheets or siltstones. They show a wide range in form and dimensions, the larger ones being 20 cm (8 inches) or more across while the smallest may be only one cm or so across (e.g. Plate 14B). Some beds have a fairly continuous pattern of flute marks covering most of the sole, but more commonly the flutes are isolated and separated by flat areas. These flat areas may show a fairly



A. Sole of bed showing longitudinal ridge and furrow structure modified by pre-existing transverse ripples. Non-directional cellular pattern developed in ripple troughs. Current from bottom to top. Scale in centimetres.



B. Small isolated flute marks on sole of sandstone bed. Note coarse-grained filling, including small pebbles; also obstacle scours behind quartz grains on flat part of sole. Scale in centimetres.

prominent lineation due to obstacle scours behind quartz grains (Plate 14B). Most of the flute marks have the characteristic deepened upcurrent end and flaring downcurrent end, but the amount of relief and the degree of spreading varies greatly. Many of the closely-spaced flutes show marked "corkscrewing" at the upcurrent end. Many have terraced sides due to differential scouring of harder and softer laminae in the underlying sediment. On some beds, flute marks occur along the sides or base of larger channel structures (e.g. Plate 15A).

The type of sediment filling the flutes varies from bed to bed and within beds. In some cases the flutes are filled or partly filled with structureless unsorted conglomeratic material which is distinctly coarser than the sediment above (Plates 9B, 10A, 14B). In these examples there is commonly a sharp break in grainsize at the top of the coarse material, corresponding to the first appearance of lamination in the bed. In other beds there is a gradual decrease in grainsize upwards through the flute. A third type has the flutes completely filled with laminated or cross-laminated medium to fine sand (e.g. Plate 13B), although lateral tracing of such beds may show a basal conglomeratic division elsewhere.

Flute marks have generally been considered as having originated from vortices or eddies which developed within the current and impinged on the bottom sediment (e.g. Dzulynski and Walton, 1965), although opinions have differed as to the orientation of these eddies with respect to the bottom. Similar structures have been recorded from alluvial deposits. More recently, Allen (1968, 1969) has argued that small irregularities on the floor are necessary to initiate flute formation, which he considered to occur because of flow separation over the irregularity and consequent over-deepening of the initial hollow. It is generally agreed that the flutes which occur on turbidites were cut by the turbulent frontal part of the current, and the experimental evidence tends to confirm this (Dzulynski and Walton, 1965). The existence of scoured intra-bed surfaces, such as previously described, would suggest, however, that several phases of flute formation may occur within the one current.



A. Sandstone filling small channel structure cut in incoherent slump sheet. Note laminated filling, also flute mark with gravelly filling at base.



B. Sole of sandstone bed showing superimposed load marks with preferred orientation.

(v) Channels

Scour structures larger than flute marks are referred to as channels, and are fairly common in those beds overlying slump sheets (e.g. Plates 11A, 15A). They vary in exposed dimensions from about 20 inches (50 cm) wide and 3 inches (8 cm) deep to about 6 feet (2 m) wide and 20 inches (50 cm) deep. Some of the channels have smaller scour structures, probably flute marks, developed beneath them. All gradations exist from channels completely filled with laminated sandstone through those with a basal layer of conglomeratic material to channels completely filled with fine conglomerate, and there are strong indications that all these variations can occur along a single bed.

Some of the irregular channel-like scour structures which occur on beds overlying coherent slump sheets (e.g. Plate 27) appear to coincide with original hollows in the pre-bed surface, suggesting that these slump sheets were sufficiently cohesive to preserve slight humps after movement had ceased.

b. Load Structures

Load structures are best developed on beds overlying siltstones or slump sheets, and are distinguished from scour marks by the lack of truncation of the laminae in the underlying sediment, by the parallelism of these laminae with the sides of the structure, and by the tendency towards concentric lamination within the structures. Many scour marks show over-deepening due to loading, however, and there appear to be all gradations from one type to the other. Many structures which appear to be due to loading show a distinct preferred orientation (e.g. Plate 15B), suggesting a primary origin from current scours.

The load structures are generally rounded or oval in plan, but may be irregular or elongated. They vary in size from one cm or so up to more than 12 inches (30 cm) across, and the depth range is similar. The underlying sediment usually forms pointed wisps (flame structures) projecting up into the bed between the load marks. The orientation of these wisps ranges from horizontal to vertical, but only rarely is there a preferred orientation of adjacent wisps.



A. Large load structures on sandstone bed overlying small slump sheet.



B. Series of load structures along sandstone overlying small slump sheet. Complex internal structure of load marks largely due to mixing with slump material.

The load marks usually form a fairly continuous pattern over the sole, and may overlap or be superimposed on one another (Plate 15B). They tend to be of fairly uniform size on any one bed, and there is at least a general correlation between the size of the structures and the thickness of the sandstone bed. The thickness of the underlying siltstone also has an obvious influence on their size.

The largest load structures are found on beds which truncate slump sheets (Plate 16A, B), suggesting that the latter were very poorly consolidated at the time. The depth of these large structures commonly exceeds the thickness of the undeformed upper part of the bed, and in a few cases the bed is actually discontinuous between them. Some of the load structures have a complex internal structure which appears to be partly due to mixing with the underlying unconsolidated material before stabilization by further deposition (e.g. Plate 16B).

Many of the load structures are filled or partly filled with sediment which is distinctly coarser than the sediment above. As with the flute marks, the upper contact of this material may be sharp or gradational. When the contact is sharp the overlying sand is usually distinctly laminated. In the example shown on Plate 16A the contact is sharp, as seen to the right of the rule, but is rather irregular as though the structure had continued to sink even during the later phases of the current. In most cases, however, the load structures are overlain by parallel-laminated undeformed sandstone, and the top of the bed is also usually flat. This strongly suggests that the formation of the structures was completed prior to deposition of the upper part of the bed, i.e. they are syn-depositional.

Small pockets filled with coarse sediment occur along the interface between two sandstone beds in some places (e.g. Plate 10B). It is unlikely that any density difference was involved in the formation of these, but their similarity to load structures suggests a comparable origin. Possibly the underlying sand was saturated and "quick", and allowed some of the rapidly-deposited coarse material to sink into it before equilibrium was established by extrusion of water and deposition of more sediment.

5. Graded Bedding

Graded bedding is characteristic of the Group B sandstones. Although statistical data have not been collected, observations suggest that at least 30% of the beds are clearly graded and of the remainder many can be shown to be graded on detailed inspection. The form of the grading varies greatly from bed to bed, and all the types shown by Dzulynski and Walton (1965, p.171) appear to be present. Ideal grading, however, in which there is a continuous and gradational decrease in grainsize from bottom to top, is fairly rare. More commonly, there are sharp breaks in grainsize within the bed (discontinuous or interrupted grading), particularly between conglomeratic material at the base and the overlying sand, and between the main sandstone part and the upper silty part. In some cases the presence of these breaks makes it difficult to decide if one bed or several beds are represented.

The coarsest material is commonly restricted to sole structures or to a thin zone at the base of the bed, while the main part of the bed may consist of fairly uniform sandstone. In such cases the lower conglomeratic material is usually unsorted and unlaminated, whereas the sandstone is better sorted and more or less distinctly laminated. Thus the break in grainsize often marks the beginning of lamination and sorting. Gradational contacts from conglomeratic material to sandstone are also common. Grading is usually not so distinct in those beds which lack both the basal conglomeratic part and the upper silty part, although careful examination usually shows a slight decrease in grain-size upwards.

Multiple grading, or grading with lenses (Crowell et al., 1966, p.18), in which one or more coarse-grained bands or lenses occur above the base of the bed, is also common. In some cases of this type it is possible that the coarse-grained layer marks the base of a separate bed, as previously discussed.

Cumulative curves of grainsizes from three thin sections taken at intervals within the one sandstone bed are given in Fig. 9 (p.92; Numbers B2(a), B2(b), B2(c)). B2(a) is from the base of the bed, B2(b) from about 4 cm above the base, and B2(c) from 10 cm above the base at the top of the division of parallel lamination. These show the

general decrease in grainsize (accompanied by an improvement in sorting), but since each thin section encompasses several laminae of slightly different grainsizes the actual complexity of the grading is not shown.

6. Internal Structures

A constant sequence of internal structures is characteristic of the Group B sandstone beds, and comprises four main divisions, viz. (a) basal division of structureless coarse material which may or may not be graded; (b) division of horizontal (parallel) lamination, mostly in medium to fine sand; (c) division of cross-lamination and ripple mark (or "convolute lamination" in rare beds); (d) upper division of laminated siltstone. Some beds do not show the complete sequence, e.g. the two upper divisions are missing from many beds in composite sequences, while the lower division is missing from many of the thinner beds. Although statistical data have not yet been collected, the most commonly-occurring sequences appear to be abcd, abc, ab, and bcd. The Group A siltstones are characterized by cd sequences. Many beds show lateral variations in the sequence, particularly as regards the presence or otherwise of the basal structureless division.

A similar sequence of internal structures is recognizable in flysch sandstones in most areas, and has been documented by Bouma (1962) and others (see review by Duff et al., 1967). The interpretation of this sequence is discussed when the mode of emplacement of the sandstones is considered.

a. Basal Structureless Division

Many beds have a basal structureless division which is coarser-grained than the overlying laminated division and is commonly conglomeratic. As well as being structureless, this division usually shows a distinct lack of sorting as compared with the laminated sandstone, and may contain the complete range of grainsizes from pebbles to fine mica. When conglomeratic, the sediment is similar in all respects to the typical Group D conglomerate. The contact between this division and the overlying laminated sandstone is commonly abrupt, and is usually accompanied by a marked improvement in sorting due to loss of the coarsest and finest grains.

The proportion of each bed occupied by this division varies greatly, and there are all gradations from the conglomerate beds of Group D, which consist almost wholly of this division, to sandstone beds with only a thin gravelly layer at the base. Similarly, there are all gradations from the thick Group C sandstones, which also consist mainly of this division (non-conglomeratic), through beds in which the structureless and laminated divisions are of equal importance, to beds in which the laminated divisions predominate. In general, the importance of this division seems to decrease with bed thickness, and thin beds in which it predominates are rare. In many beds, this division is restricted to bottom depressions such as current scour marks and load structures, while in other cases it occurs only as isolated lenses or even as isolated pebbles (e.g. Bed A of Fig. 8).

Interpretation

The interpretation of the basal structureless division is considered to be fundamental in reconstructing the nature of the depositing currents, since this division was deposited first and hence most closely reflects the conditions within the current during transport. It is apparent that the currents must have been capable of carrying very coarse sediment, including large pebbles in some cases, mixed with all the finer grades, and of depositing this mixture without sorting it or producing lamination (except in the later stages). The topographic control of deposition, as indicated by the preferential filling of bottom depressions, indicates that the currents were bottom-hugging. The presence of coarse-grained depression-fillings, as noted by Sanders (1965), also indicates that the initial gravel-bearing phase of the current was capable of passing over an area without depositing sediment (except in the depressions). These features and others are discussed more fully when the mode of emplacement of the sandstones is considered.

b. Division of Parallel Lamination

The basal structureless division is normally overlain by a division of parallel lamination. Many of the thinner beds show parallel lamination throughout most of their thickness, and even deep sole

structures may be laminated throughout (e.g. Plate 11A). The prominence of the lamination varies greatly according to the degree of weathering and the differences in grainsize and composition between adjacent laminae. Some beds which appear structureless are found to be laminated on very close inspection. The base of the laminated division may be gradational but more commonly there is an abrupt contact with the underlying structureless division.

The thickness of individual laminae varies from one or two grain diameters up to several mm or more, and most appear to be more than one grain diameter in thickness. The difference in grainsize, particularly maximum grainsize, between adjacent laminae may be very considerable, e.g. from fine sand to granules.

Parting lineation (Crowell 1955) is associated with the parallel lamination in most beds. This structure may be of the "parting-plane lineation" type (McBride and Yeakel, 1963), which consists of minute grooves and ridges developed on a single lamination plane, or the "parting-step lineation" (*ibid*) formed by linear breaks between adjacent laminae. Since both types always seem to occur together, the differentiation of terms is probably unnecessary. As shown by McBride and Yeakel, the lineation reflects the orientation of the long axes of the quartz grains and is hence an excellent current direction indicator.

In some beds a series of lamination planes with parting lineation are exposed in three dimensions, making it possible to determine the current direction at several points within the bed. While these usually remain parallel, there are a few cases where significant changes in current direction are recorded. In one bed, the lineations on successive laminae indicate a clockwise swing in current direction of 25° through a vertical thickness of 2 cm of sediment.

Origin of parallel lamination

The origin of parallel lamination in sands, and in "turbidite" sands in particular, has long been a problem. Most authors have attributed it to pulsations in the flow, but the work of Moss (1962, 1963) and Kuenen (1966) suggests that it may be the natural result of traction processes on the bottom and be largely independent

of current pulsations. According to these authors there is a strong tendency for grains of similar dimensions to congregate into flat sheets from which dissimilar grains are rejected. The overlapping and superposition of these sheets as they come to rest is thought to result in parallel lamination.

Moss' conclusions were based mainly on the physical analysis of naturally-occurring clean sands and pebbly sands (i.e. not turbidites), while Kuenen's work involved the experimental formation of lamination from suspensions in a circular flume. Whether these results can be directly applied to turbidite lamination has not yet been confirmed. The influence of eddies and fluctuations during deposition from turbidity currents, particularly as regards the formation of thick laminae and abnormally fine-grained or coarse-grained laminae, must be considered. The fact that most turbidite laminae appear to be at least several grain diameters in thickness may also be significant.

The association of parting lineation (primary current lineation) with the parallel lamination has generally been considered as indicating upper flow regime conditions during deposition, following the work of Allen (1964). According to Allen, this structure only forms during the plane bed (with sediment movement) phase which occurs in the lower part of the upper flow regime and in the transition to the lower flow regime (e.g. Simons et al., 1965). The significance of this is discussed further when the mode of emplacement of the sandstones is considered.

c. Cross-Lamination and Ripple Mark

Cross-lamination is developed in the upper parts of many of the sandstone beds. In some of the thinner beds it extends right to the base, and even fills the sole marks in some cases (Plate 13B). The form of the cross-lamination varies, but most examples are of the small-scale trough type. Individual foresets are generally asymptotic or parallel to the base, suggesting up-building of the ripples as well as forward progression.

Some sandstone layers have sharp tops with well-defined ripple marks but show few if any cross-laminae. The ripple troughs

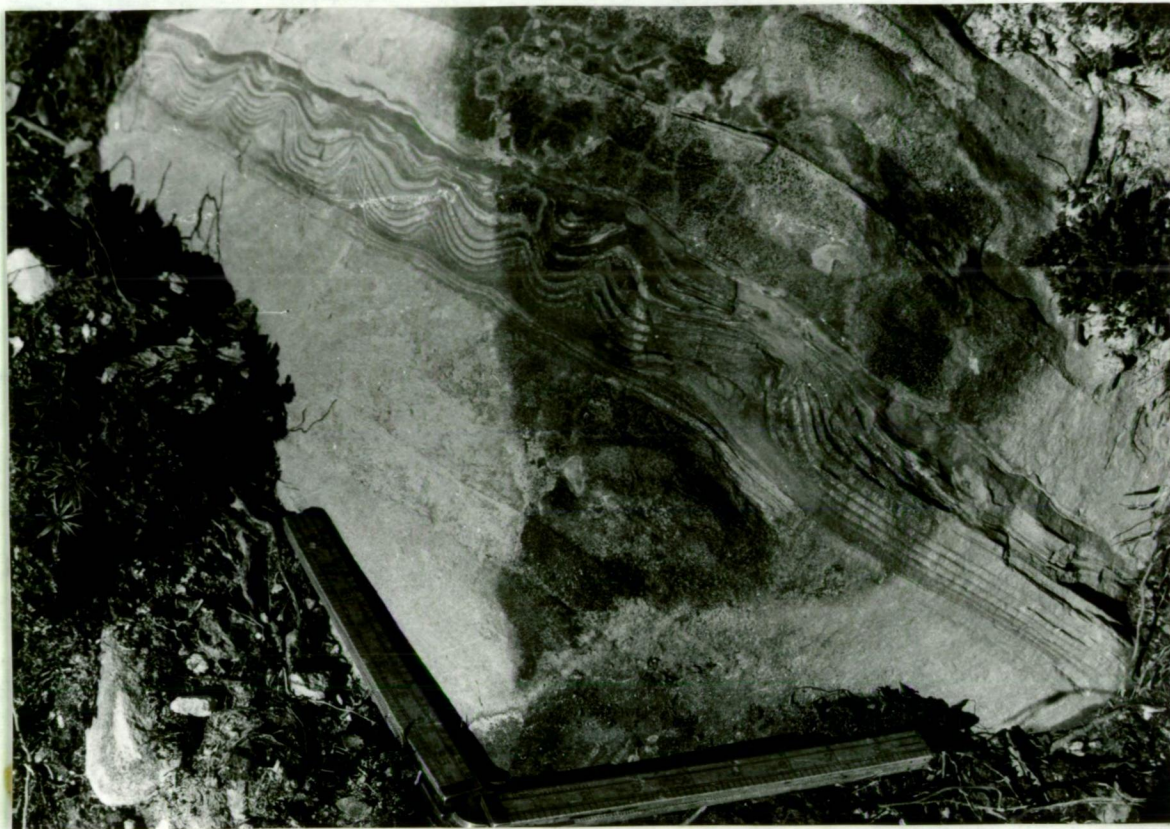
erode deeply into the underlying parallel laminae, and the ripples are usually overlain abruptly by siltstone. Other sandstone layers of similar type show rapid but gradational contacts with the overlying siltstone, however, suggesting that the absence of cross-laminae in the sharp-topped layers is a syn-depositional phenomenon rather than the result of reworking by a later non-depositing current.

Well-preserved ripple marks are seldom exposed in plan view. Those that can be seen are usually rather irregular (non-linear) in form, with rounded, slightly asymmetrical crests enclosing oval or sub-circular troughs. The cross-laminae dip into the troughs from three sides and in some cases from all sides, i.e. the troughs are aggradation-al. The ripple crests on some beds are mostly concave downcurrent, approaching the crescentic or lunate type, and many beds show the "rib-and-furrow" structure thought to be formed by ripples of this type (Dzulynski and Walton, 1965, p.175). On other beds, however, the crests are mainly convex downcurrent and hence approach the linguoid type. Straight-crested ripple marks are rare. Ripple amplitudes range from about 4 mm to 2 cm, while wavelengths range from about 3.5 cm to 15 cm.

d. Deformed Lamination

Deformed lamination, or "convolute lamination", is very rare in the graded beds of the Singing Creek Formation. Most of the sandstone layers have fairly sharp upper contacts with the overlying siltstone or with the next sandstone bed, and the intermediate zone of inter-laminated fine sand and silt, which is most prone to deformation, is usually absent. One beautifully exposed zone of deformed lamination is shown in Plate 17.

In this example, the folds are developed in a 6 cm (2.4 inches) thick division of alternating fine sandstone and micaceous siltstone laminae forming the top of a graded bed. The sequence is truncated by the base of the overlying graded sandstone. The folds are fairly regular and have an average wavelength of 8 cm (3.25 inches). They are apparently arranged in a linear parallel fashion, since a joint surface at right angles to that which exposes the folds shows only flat parallel laminae (Plate 17A) for a distance of 30 cm (12 inches).



A. Outcrop showing folded lamination in upper silty part of graded sandstone bed. Surface at right is parallel to fold axes. Several graded beds present. Scale in inches.



B. Detail of folded lamination. Sandy laminae are light-coloured. Note basal decollement zone and thinning of the basal sandy laminae beneath synclines; intrusive nature of anticlines and thinning of most laminae over the anticlines; upper decollement zone and fine cross-lamination at top. Current from right to left. Wavelength of folds 8 cm.

Slight indentations along this latter surface show the laminae turning up towards the crest of the first anticline.

The folds consist essentially of sharp, partly intrusive anticlines separated by broad synclines. That the anticlines were growing during deposition is indicated by the thickening of most of the sand laminae in the synclines. Several of the basal sand laminae show very marked thinning beneath the synclines, however, and concomitant thickening in the anticlines (Plate 17B), strongly suggesting that there was migration of sediment from beneath the synclines to "feed" the growing anticlines. Intrusion of the anticlinal sediment into the overlying laminae is shown in several places. In some of these, the intrusive material was derived from a lamina in the middle or upper part of the sequence as well as from the more deformed basal laminae. The pattern becomes less regular towards the top as the folds are progressively smoothed out, and several subsidiary folds are developed in this part. Very fine-scale cross-laminae have been deposited down-current from most of the anticlines in this upper part, and the dip of these becomes flatter as the synclines are filled. The current direction indicated by the cross-laminae agrees with that indicated by the slight overturning of the anticlines, and was apparently perpendicular to the axes of the folds.

Interpretation

It is apparent that the folds developed as sedimentation was occurring, and that the structure was completed before deposition of the bed had ceased. There is no evidence to indicate that the folds were initiated by current ripples, since the only cross-lamination present is near the top of the sequence. The initiation of the folds is best explained in the manner suggested by Sanders (1960), in which a lamina (siltstone in this case) becomes too cohesive to allow formation of current ripples and instead is dragged into a series of anticlines. These grow by a "decollement-type of adjustment" and thus may be quite independent of the lamina below. Two decollement zones, both consisting of siltstone, are evident in the present example, one at the base and one near the top. The suggestion by Sanders (*ibid*, p.416) that the

anticlines grow "by lengthening of the deformed layer and by flow of material from both sides towards anticlinal crests" finds elegant confirmation in this example.

Thus the folds are probably analogous to current ripples in that they represent bottom roughness (form-drag) features developed in response to the shearing stress imposed by the current (Bagnold, 1954, 1956), the difference being that the initially-deposited silt was too cohesive to form true ripples. A very similar kind of intra-stratal folding may result from the upward growth (and downcurrent overturning) of ripple crests during sedimentation (e.g. Kuenen, 1953; Dzulynski and Walton, 1965).

e. Upper Division of Laminated Siltstone

While many of the sandstone layers are overlain directly by another sandstone, there are some which pass upwards into siltstone. The contact with the siltstone may be gradational but is more commonly abrupt, so that it is not always obvious that the silt was deposited by the same current as the sand. The siltstone is similar to that described for Group A, consisting of dark micaceous laminae alternating with lighter-coloured quartzose laminae. Micro-grading of the coarser laminae is common. Where a sandstone layer is isolated within siltstones it is usually difficult to decide how much of the overlying siltstone belongs to the same depositional unit as the sandstone.

The absence of this division from most beds in composite sequences is probably due to both non-deposition (due to excess velocity in the tail of the current) and to erosion by the following current.

7. Composition and Provenance

About 35 thin sections of Group B sandstones have been studied (see Appendix B for numbers). They show a striking uniformity of composition, paralleling that of the siltstones, with only three major components, viz. quartz (and quartzite) grains, micaceous minerals, and carbonate. Authigenic quartz is present to some extent in most slides. Rock fragments apart from quartzite are rare in most samples, but include quartz-sericite-schist, quartz-muscovite-schist, chloritic schist, slate, dolomite and mudstone (probably intraformational). A few grains of plagioclase are present in some slides, but

igneous rock fragments are absent. Tourmaline is the most abundant accessory mineral, the other main ones being zircon and topaz. Fossil fragments (trilobites and brachiopods) are found in some beds and appear to have been transported.

a. Modal Analyses

Modal analyses of six thin sections of Group B sandstones, including one from the basal conglomeratic division of a bed (No. B1), as well as four thin sections of Group D conglomerates and two thin sections of Group C sandstones, are given in Table 1. Cumulative curves of the grainsize distributions of these same rocks are shown in Fig. 9, and are discussed under Texture. The analyses were done with a Swift point counter, using a grid spacing of 0.3 mm, and about 1500 points were counted in each case, covering most of the thin section. Grainsize analysis of the quartz grains was combined with the modal analysis, such that each size division (one phi unit) was counted as a separate mode.

It was found impracticable to attempt to distinguish between clear quartz, undulatory quartz and polycrystalline quartz in the detrital fragments, because there are many grains which are intermediate in type between these. As pointed out by Blatt and Christie (1963) the observed degree of undulatory extinction depends largely on the orientation of the grain in the thin section, and the amount of rotation necessary to extinguish all the grain may range from zero to 90° depending on this orientation. Hence all the quartzose grains are grouped together. An attempt was made to distinguish authigenic quartz from detrital quartz in a few cases, but this was not possible because the contacts between the two types were seldom recognizable (rough counts on several of the sandstones suggest 5-10% authigenic quartz). Since much or all of the authigenic quartz was probably derived from pressure solution between grains, the combination of the two probably reflects the original detrital quartz content.

It is obvious that such a small number of analyses will not be representative of the sandstones, but at least they serve as an indication of the compositions present and permit more accurate

TABLE 1. MODAL ANALYSES OF SANDSTONES AND CONGLOMERATES

Locality		Description	Ref. no. in Fig.9	Quartz grains	Carbonate	Muscovite	Sericite + chlorite	Total mica	Other rock fragments	Remarks	Thin section number
GROUP B SANDSTONES	West slope Flagstone Knoll	Conglomeratic part of Bed B2 in Fig.8	B1	85.6	6.6	0.7	2.1	2.8	5.0*	* Mainly carbonate rock fragments	37250
	"	Single graded) base bed with) 4 cm flute marks) 10 cm	B2(a)	92.7	1.1	2.1	3.6	5.7	0.5) Upward decrease in quartz, increase in carbonate	37257
			B2(b)	88.1	8.3	1.1	2.5	3.6	-		37257
			B2(c)	81.0	10.9	4.2	3.9	8.1	-		37258
	"	Part of compositely bedded sequence shown in Plate 7B	B3	82.6	10.8	1.4	5.2	6.6	-		37256
	Cliff near head of Singing Creek	Thin rippled bed with load casts	B4	86.9	3.0	1.7	8.1*	9.8	0.3	* Sericite introduced from flame structures	37259
GROUP C SANDSTONES	Crest of Flagstone Knoll	Part of bed with dish structure	C1	86.7	-	7.7*	3.3	11.0	2.3	** High muscovite content	37264
	"	Sandstone with elutriation columns	C2	85.2	-	8.8*	5.6	14.4	0.4		37265
GROUP D CONGLOMERATES	South slope Flagstone Knoll	Lower part of Bed 1 in Fig.10	D1(a)	93.2	-	0.3	6.5	6.8	-	* Schistose rock fragments	37275
	"	Repeat of D1(a)	D1(b)	92.2	-	0.2	6.3	6.5	1.3		37275
	Ridge above Boyes River	Graded fine conglomerate bed	D2	90.3	-	0.6	2.4	3.0	6.7*		37276
	South slope Flagstone Knoll	Base of Bed 3 in Fig.10	D3	92.8	2.3	2.4	2.5	4.9	-		37274
	"	Sandstone part of Bed 2 in Fig.10	D4	83.5	10.1	3.1	2.9	6.0	0.4		37277

See Fig. 9 for grainsize distributions of these samples.

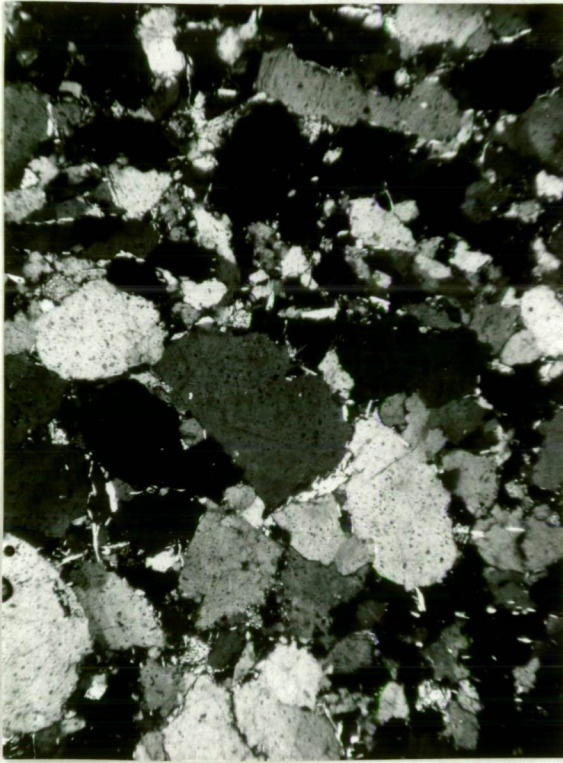
estimates of the composition of other samples by comparison. The high quartz content (80-95%), variable carbonate content (0-12%), and variable but generally low mica ("matrix") content (3-10%) of the sandstones are apparent in the analyses. Any future study of this type should take into account the variations within individual beds, since these may be as great as or greater than the variations between beds.

b. Compositional Components

Quartzose grains constitute at least 80% of virtually every sample. A high proportion (at least 60%) of the grains coarser than 0.25 mm are either undulatory or polycrystalline. Some contain scattered small mica flakes and a few large grains have rows of small oriented tourmaline crystals. Thus there is little doubt as to the metamorphic origin of the grains. The ratio of clear grains to undulatory grains increases as grain size decreases, as might be expected. Contacts between adjacent grains may be sharp and straight, sutured and interlocking, or vague and poorly defined (Plate 18A). In many cases the contacts are marked by a thin discontinuous line of sericite flakes, suggesting partial recrystallization or replacement of original interstitial material. Clearly defined secondary overgrowths of quartz are generally not apparent, although the many sutured contacts in those samples or laminae lacking interstitial material indicate a significant amount of secondary quartz.

Micaceous material constitutes 5-10% of most samples, but the proportion varies greatly both within and between beds. Those beds which grade upwards into siltstone show an obvious increase in mica content upwards. Muscovite and sericite are the predominant types, with only minor amounts of biotite and chlorite. The muscovite and biotite appear to be almost wholly detrital, since many of the flakes show weathered edges and are bent between adjacent quartz grains. The flakes are commonly aligned parallel to bedding, particularly in laminated beds. The sericite occurs as fine aggregates and flakes sandwiched between quartz grains (Plate 18A), but only rarely is there sufficient to form a continuous "matrix". How much of this material is detrital and how much authigenic is difficult to estimate, but its

- A. Photomicrograph of typical Group B sandstone. Note predominance of quartz grains; lack of continuous matrix; sericite flakes between many grains; mixture of angular and well-rounded grains. Crossed nichols. Width of field is 1 mm. Specimen is No. B3 of Table 1 and Fig. 9 (No. 37256 of Dep't. collection).
- B. Photomicrograph showing introduction of dark sericitic matrix from flame structure into base of sandstone bed. Normal mica-poor sandstone can be seen towards right and left margins. Ordinary light. Width of field is 2 mm. Specimen is No. B4 in Table 1 and Fig. 9 (No. 37259 of Dep't. collection).
- C. Conglomeratic part of sandstone bed. Note bimodal nature with small pebbles and abundant sandy matrix; similarity to Group D conglomerate of 18D. Darker blebs are carbonate. Ordinary light. Width of field is 2 mm. Specimen is No. B1 of Table 1 and Fig. 9, and is from Bed B1 of Fig. 8 (No. 37250 of Dep't. collection).
- D. Typical fine conglomerate of Group D. Note bimodal nature, with micaceous material probably forming third mode; mixture of sub-angular and well-rounded fragments; rows of small tourmaline crystals in large quartzite fragment at centre. Ordinary light. Width of field 2 mm. Specimen is No. D1 of Table 1 and Fig. 9, and is from Bed 1 of Fig. 10 (No. 37275 of Dep't. collection).



A.



B.



C.



D.

abundance in the siltstones suggests it could be largely detrital. Chlorite occurs as detrital grains, as an alteration product of the other micas, and as a fine intergrowth with sericite in some cases, but it is generally a very minor constituent of the sandstones.

The carbonate content varies greatly, from virtually zero in some beds to more than 30% in rare cases, the average being about 10%. It occurs mainly as crystalline blebs of various sizes located between the quartz grains and partly replacing them. Many of the blebs show limonite-stained edges, suggesting the carbonate is an iron-rich variety. Detrital grains of similar-looking or finer-grained carbonate occur in a few of the sandstones and conglomerates and are possibly intra-formational. The carbonate is very susceptible to weathering, and its original presence in many samples is indicated only by small limonite-stained cavities.

Tourmaline is a very common accessory in nearly all slides, forming small, roughly equidimensional grains up to 0.2 mm diameter. Brown and green varieties are by far the most predominant, and occur in about equal proportions. Blue tourmaline also occurs, but is very rare. Most of the grains are somewhat rounded, but angular and euhedral grains also occur. Some grains show evidence of considerable secondary growth around cores of slightly different colour, while others contain large clear inclusions, apparently of quartz. The presence in some Group D conglomerates of quartzite rock fragments containing abundant tourmaline crystals (e.g. Plate 18D) suggests that most of the tourmaline is of metamorphic origin. The predominance of brown and green varieties also suggests this.

Small, well-rounded zircon grains up to 0.1 mm are present in most slides, while rounded topaz grains also occur but are less common.

Nearly all beds show vertical variations in composition, particularly in the relative proportions of quartz and mica. Beds which are graded from laminated sandstone at the base to siltstone at the top show an obvious overall increase in mica content upwards. Those beds which have a basal coarse-grained structureless division, however, may also show an increase in mica downwards, since the basal

division is commonly enriched in fines with respect to the division of horizontal lamination. Bed B2 in Table 1 shows this to some extent. As suggested by Stanley (1963), the vertical variations within beds may be as great as any of the inter-bed variations. Concentration of mica in some laminae and not in others also complicates the picture.

c. Provenance

The virtual absence of feldspar and igneous rock fragments, combined with the abundance of quartzite rock fragments and the presence of muscovite, schistose rock fragments and metamorphic tourmaline, all indicate a low-grade metamorphic provenance. Palaeogeographic evidence, including palaeocurrent analysis, indicates that the source area for the sediments was the Tyennan Geanticline immediately to the west. This is a basement structure of Precambrian metasediments, dominated by quartzites, quartz schists and quartz-mica schists, from which all of the detrital material could have been derived.

8. Texture

The present author follows Folk (1961) and Dott (1964) in considering textural maturity in terms of three factors, viz. percent of "clay" matrix, sorting of the sand fraction, and rounding.

a. Matrix

Most authors concerned with the description of turbidites have regarded the material finer than about 0.03 mm (5 phi) as the "matrix", partly because of the difficulty of identifying grains finer than this. However, the term "matrix" should imply a distinctly finer-grained portion within which the coarse grains are contained, and hence, as pointed out by Kuenen (1966, p.268) a conglomerate may have a "matrix" of entirely sand-grade material. Using the term in this sense it is apparent that most of the sandstones have very little matrix, whereas the conglomerates and the conglomeratic division at the base of some sandstone beds, have an abundant sandy matrix (e.g. Plate 18C, D).

Most of the sandstones consist very largely of quartz grains of all sizes down to about coarse silt grade, with most of the

grains in contact and most of the interstices filled with either carbonate or secondary quartz. The micaceous minerals are the only constituents in these rocks which can be described as matrix, but the total amount of these seldom exceeds 10% and in many cases is less than 5%. While it is probable that some original matrix has recrystallized or has been replaced by secondary quartz or carbonate, this could only add a few more percent. Very few of the sandstones can have had a continuous fine-grained matrix such as characterizes many "greywacke" turbidites. In this respect, then, the sandstones are moderately immature to submature in the sense of Folk (1961) and Dott (1964).

In one sandstone bed with deep load casts fine micaceous matrix has been introduced throughout the lower part of the bed by way of flame structures which protrude up from the underlying siltstone (Plate 18B; No. B4 of Table 1). The penetration of this sericitic material between the sand grains for at least several cm beyond the ends of the flame structures indicates that it was highly mobile and possibly suggests fluidization due to the weight of the suddenly-emplaced sand.

Interpretation of matrix content

The origin and significance of the matrix in turbidites have been much discussed. Both Cummins (1962) and Kuenen (1966) have argued that the abundant matrix (15% or more) which is characteristic of many "greywacke" turbidites must be due to post-depositional alteration of unstable minerals. Part of Cummins' argument involved the decrease in the proportion of greywackes, with respect to other types of flysch sandstones, from Precambrian to Recent, and the scarcity of matrix in experimental and modern deep-sea turbidites. Kuenen argued from experimental and theoretical evidence that the lutum content retained in a turbidite sand reflected the lutum content of the original suspension, and should not exceed 10%. Dzulynski and Walton (1965, p.29) have pointed out, however, that there is only a weak correlation between matrix content and the proportion of labile constituents in the Ordovician greywackes of Scotland, and suggested that the amount of fine material in the source sediment would be important.

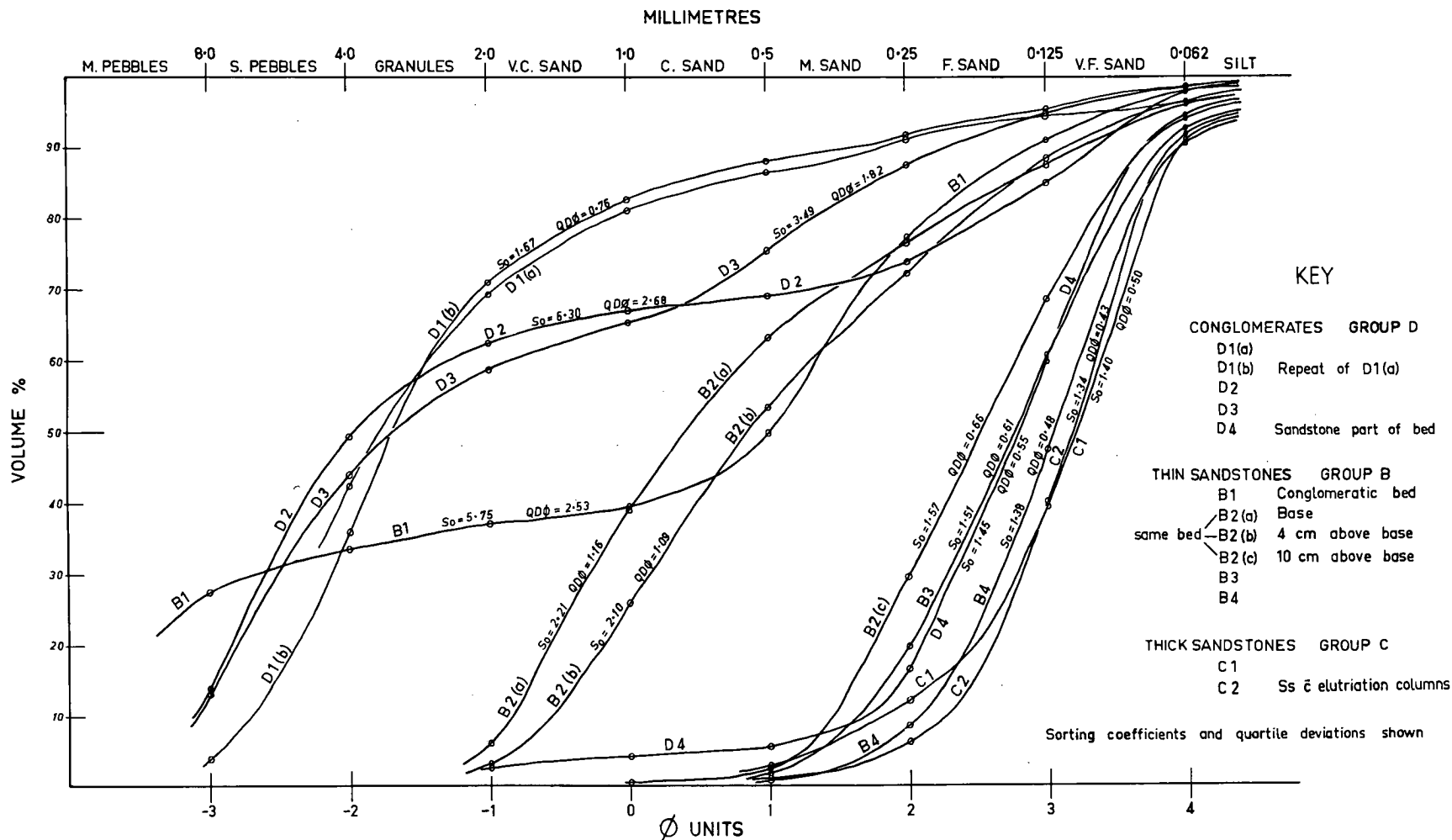
The experience of the present author suggests that both metamorphism and nature of the source sediment, as well as a third factor, viz. the stage reached by the current when deposition occurs, were probably important. The shearing which has obviously affected many geosynclinal turbidites, and the consequent inter-granular movement, has probably been as effective as load metamorphism (i.e. pressure solution between grains) in adding material to the matrix. In the case of the Group B sandstones the low matrix content seems to be partly due to a low proportion of fines in the source material, and, to some extent at least, to the tractional movement which accompanied deposition of the greater part of most of the Group B beds and which produced the almost ubiquitous lamination. Most of the thin sections examined were from the division of parallel lamination (b) which forms the bulk of most of the thin sandstone beds. As discussed later, the deposition of this part was probably accomplished by the middle part of the current, in which tractional and sorting processes were most pronounced. The underlying structureless division (a) and the overlying silty division (d) tend to have a higher matrix content (e.g. B2 in Table 1). The effect of load metamorphism, in contrast to Cummins' suggestion, has apparently been to reduce the matrix content in favour of secondary quartz.

b. Grainsize Analysis and Sorting of Sand Fraction

(i) Method

Cumulative grainsize curves of the samples used for modal analysis (Table 1) are shown in Fig. 9. Because of the large grain-sizes involved with the conglomerates, and the lack of suitable polished slabs or large thin sections, it was decided to use the technique of modal analysis, rather than grain counts, to determine the proportions of the various grainsizes present. Thus each grade (one phi unit) on the Wentworth scale was regarded as a mode, and the areas occupied by each were measured as part of the overall modal analysis of the slide. This was done by first measuring the maximum diameter of the grain with a micrometer eyepiece, to give the appropriate size grade, and then counting the number of intersection points across the grain.

FIGURE 9



SIZE DISTRIBUTIONS OF QUARTZ GRAINS IN SANDSTONES AND CONGLOMERATES

DATA FROM THIN SECTION ANALYSIS

The results can thus be expressed as volume or weight percent rather than number percent as is usual with thin section size analysis.

About 1500 counts at a grid spacing of 0.3 mm (the smallest possible with the equipment available) were made in each case, covering most of the thin section. This meant that at least 500 quartz grains were measured in each section. All grains finer than 0.06 mm were grouped together as a single silt mode, since accurate measurements at this level were not possible. Hence the graphs are unlikely to be accurate beyond the very fine sand grade. The sections used were cut normal to bedding in most cases, but this was not possible with the conglomerates since the samples were mostly not oriented. To test consistency the first section measured (D1(a) in Fig. 9) was repeated after the last section was done, and the result suggested that a reasonable level of consistency had been maintained. No correction for sectioning effect has been made.

A sample of thin section size will obviously not be representative of a conglomerate, since only a few of the larger clasts may be present. Thus the method tends to under-estimate the proportion of the larger sizes to some extent. The size distributions obtained are similar to those suggested by visual examination of the conglomerates, however, and show the obvious bimodal nature of these rocks. Similarly, the main features of the grainsize distributions of the sandstones, which are fairly obvious in hand specimens and thin sections anyway, appear to be reproduced in the graphs. Because of the few data, however, the following generalizations must be regarded as somewhat tentative.

(ii) Sorting

Although the sandstones show a considerable range in sorting, it is apparent that most fall within Trask's well-sorted class (S_o less than 2.5), and some are very well sorted. The sorting coefficient S_o ($= \sqrt{Q_3/Q_1}$) is not the best indicator of sorting in these rocks, since some samples (e.g. D1, D4) have "tails" of coarse or fine sediment which, because they constitute less than 25%, do not affect the value of S_o . Other coefficients have not been calculated, however, since the graphs are unlikely to be accurate at the high and low values,

and the important relative variations are clear enough from the shape of the curves. The conglomerates and the conglomeratic division of the sandstone bed (B1) are obviously poorly sorted and markedly bimodal in the coarser grades, due to the presence of an abundant sandy matrix (e.g. Plate 18C, D). The silt-grade material, combined with the micaceous material, probably forms a third mode in the conglomerates and a second mode in the sandstones.

The results suggest that sorting improves as grainsize (particularly maximum grainsize) decreases, and in general this seems to be the case. Certainly, those beds which could be sampled at several points show an improvement in sorting upwards as grainsize decreases (e.g. B2(a), (b), (c) in Fig. 9). In a number of beds there is an abrupt change from a basal, poorly-sorted conglomeratic division, with an abundant sandy matrix (e.g. Plate 18C), to a laminated well-sorted division lacking conglomeratic material. This improvement in sorting is sometimes enhanced by almost complete removal of the micaceous fines. Those beds which show a gradual upward increase in the prominence of lamination also show a gradual improvement in sorting, and the correlation between sorting and prominence of lamination is apparently a strong one. If only single laminae could be measured, most of the sandstones would probably show better sorting than is indicated in Fig. 9.

(iii) Significance of sorting

According to the system of Folk (1951, p.129) the sandstones are texturally immature to submature, the boundary between the two being placed at 5% "clay" (10% according to Dott, 1964). The boundary between the submature and mature stages is placed at $S_0 = 1.3$. It is apparent, however, that the textural maturity, as measured by the sorting coefficient, will vary considerably within a single bed, e.g. many beds have a basal conglomeratic division and hence will show the complete range of values from those typical of the conglomerates (S_0 3-6) to those typical of the well-sorted fine sandstones (S_0 1.3-1.5).

Other results concerning the variation of sorting within single beds are presented by Dzulynski and Walton (1965), Stanley (1963),

and Okada (1966), and all suggest considerable internal variations but with a tendency for sorting to improve upwards. Thus the use of Folk's sorting criteria must be qualified in some respect to allow for such internal variation. For rocks showing "textural inversion", e.g. abundant matrix but a high degree of rounding, Folk advocates the use of the lowest stage of textural maturity represented, and this should also be the approach to the problem of intra-bed variability. The superimposed lateral variability of many beds, particularly those where the basal conglomeratic division is preserved only as isolated patches, increases the difficulty in this respect and makes careful observation and sampling imperative.

A significantly different and apparently more rational approach to the problem of sorting and grainsize distribution has been advocated by Spencer (1963). His two major propositions are firstly that most naturally-occurring sediments are mixtures of three or less fundamental populations, viz. "gravel" (median of -3.5 to -2 phi), "sand" (median of 1.5 to 4 phi) and "clay" (median of 7 to 9 phi), and hence are usually grains to matrix situations; and secondly that sorting may be recognized only by the degree of truncation of the original populations or mixtures. The "coefficient of sorting", as determined from quartiles, should be referred to as dispersion, and in most cases is just a measure of the degree of mixing of the populations. The actual effect of sorting (i.e. the physical process) is to reduce the dispersion and amount of one of the components until a pure fundamental population is attained. Spencer also suggests that there is a fundamental cutoff between "sand" and "clay" at about 5 phi (0.03 mm), and that silt as a genetic class probably does not exist.

The present results, although admittedly tentative appear to support Spencer's hypotheses. Although the "clay" grade is poorly represented in the analyses, it is apparent that three populations are present. The conglomerates contain all three, although the "gravel" (greater than -1 phi) and "sand" (-1 to +4 phi) predominate, e.g. No. D1 in Fig. 9 has about 70% "gravel", 20% "sand" and 10% "clay", while No. B1 has about 30% "gravel", 65% "sand" and 5% "clay". The sandstones contain only the "sand" and "clay" populations, with the

former predominating, while the siltstones consist predominantly of "clay" with variable admixtures of "sand". The "clay" population in the present case consists mainly of micaceous minerals. The decrease in grain size thus reflects a progressive loss of the coarser population, as is indicated in the cumulative curves. That the three populations probably have genetic significance is suggested by the marked differentiation into conglomerate, sandstone and "siltstone" which is characteristic of the formation.

(iii) Rounding

Roundness studies have not yet been done. The quartz grains include angular and very well rounded types (e.g. Plate 18A, B, C), and the variability in roundness is a characteristic feature of the sandstones. Many of the polycrystalline grains are elongated, tabular and sub-angular, but visual estimates indicate a surprisingly high proportion (30-40%) of rounded to well-rounded grains, particularly in the coarse sandstones. Sphericity also varies from low to high.

The high proportion of rounded grains in otherwise fairly immature sediments is a case of "textural inversion" (Folk, 1951), and strongly suggests that the rounding is inherited from a previous sedimentary cycle. This further supports the contention of a low-grade metasediment provenance, as discussed previously.

9. Classification of the Sandstones

The problems of sandstone classification are well known, and a discussion of the various classifications is beyond the scope of this thesis. However, as proposed by Klein (1963), authors dealing with the petrology of sandstones should indicate which classification they have adapted and for what reasons. The sandstones of the Singing Creek Formation present considerable problems because of the combination of apparent mineralogical maturity with textural immaturity.

The mineralogical maturity appears to be a function of the provenance rather than diastrophism, since the sediments were fairly obviously deposited under unstable conditions. The similarity in composition of these sandstones with the texturally mature sandstones of the overlying formation (Great Dome Sandstone) also implies that

diastrophism has not been critical. Hence, as suggested by Klein (1963), those classifications linking composition with diastrophism (e.g. Krynine, 1948) are not suitable. Similarly the composition is largely independent of texture, and hence those classifications linking these two (e.g. Pettijohn, 1957) cannot be applied. This is also in agreement with Klein's conclusions, although the author does not agree with Klein's arguments regarding the lack of significance of the matrix content.

An attempt was made to use the classification based on sedimentary structures as proposed by Crook (1960) and Packham (1954). This was abandoned partly because it restricted the term "sandstone" to a particular type, and hence precluded its use as a general term and in some formation names, and partly because it was not possible in some cases (e.g. the thin sandstones within the Group A siltstones, and the thick structureless beds of Group C) to distinguish with certainty between "turbidity current" and "traction current" deposits. This conclusion also accords with that of Klein (1963).

The two dominant features of these sandstones are their mineralogy and their texture, and a classification was required which treated these as independent variables. Such classifications have been proposed by Folk (1954) and van Andel (1958). Folk's classification is difficult to apply to these rocks because it requires differentiation between quartz (plus chert) and metamorphic rock fragments (including quartzite), and as previously mentioned it was found impracticable to do this. If all the grains are regarded as rock fragments, then the rocks classify as greywackes on Folk's diagram. If, on the other hand, the clear grains are differentiated as quartz, then some of the rocks, and particularly the finer-grained ones, become subgreywackes. The classification of van Andel employed the textural stages of Folk (1951) but his mineralogical classification differed in grouping quartzite fragments with quartz, while chert was grouped, rather surprisingly, at the rock fragments pole. The present sandstones would be mainly subgreywackes according to this system.

Both the above systems classify the rocks primarily on their composition as either greywackes or subgreywackes, and this agrees with the general conception that greywackes are predominantly of

turbidity current origin. However, neither system clearly differentiates these rocks from the mineralogically similar but texturally more mature sandstones of the Great Dome Sandstone. The latter, although obviously of traction current (shallow water) origin, would also classify as grey-wackes or subgreywackes because of the high proportion of quartzite rock fragments. Hence, a classification which places more emphasis on textural maturity, but which treats it as independent of composition, would seem to be ideal. Dott (1964) has proposed such a classification, based largely on the two-fold textural classification of Gilbert (1954). This system has a primary subdivision into "wackes", which are poorly sorted and have more than 10% matrix, and "arenites", which have less than 10% matrix and are generally well sorted. Subdivision of these two types is then made according to composition, using a ternary diagram with stable grains (quartz, chert, quartzite), feldspar, and unstable rock fragments as end members. This subdivision avoids confusion with respect to quartzose grains, and the present rocks can thus be classified as mainly quartz wackes.

The chief disadvantage with this system is the use of the 10% fine matrix as the cut-off between the two groups. Most of the present sandstones, although otherwise texturally immature or sub-mature, have less than 10% "clay". This is presumably the result of several factors such as lack of fines in the source sediments, sorting associated with deposition, and replacement of some of the matrix by secondary minerals during diagenesis and metamorphism. It is suggested, therefore, that a combination of textural criteria, involving both matrix content and sorting of the larger grains, be used to distinguish wackes from arenites. Such a combination was implied by Gilbert, but because most wackes have an abundant matrix, the use of percent matrix has been easiest to apply (see Dott, 1964). Unfortunately, the quantification of an alternative criterion, such as sorting of the grains, is difficult without grainsize measurements. Dott (1964) also discussed this problem, and suggested that the methods of visual estimation of sorting proposed by Swann et al. (1959) and Emrich and Wobber (1963) could be useful in this respect.

It seems likely that there must be some overlap in textural characteristics between rocks of the wacke and arenite suites (e.g. arenites rich in mica), and hence there is some danger in rigid application of quantitative values for classification. Rocks intermediate between the two groups could perhaps be classified into one or the other group according to the sedimentary structures they display, in the manner proposed by Crook (1960) for distinguishing "greywackes" from "sandstones".

10. Mode of Emplacement of the Sandstones

a. Background

The origin of flysch-type graded beds from "turbidity currents" is now generally accepted, following the original hypothesis of Kuenen and Migliorini (1950). Other theories for the origin of such beds, e.g. normal ocean currents (Hubert, 1964, 1966; Klein, 1966; Scott, 1966), "tectonic sedimentation" (Kingma, 1958, 1960), and liquefaction (Terzhagi, 1956), have not received general support, although the adherents still regard the question as controversial (e.g. van der Lingen, 1969). None of these alternatives satisfactorily explains the many unique features of flysch sandstones, as has been pointed out by Kuenen (1967, 1964) and others in numerous papers. The persistence of the argument, however, demonstrates the lack of knowledge of deep-sea processes and the difficulty of simulating these processes experimentally.

Much of the controversy seems to have arisen because of preconceptions about the nature of "turbidity currents". Instead of working backwards from the deposits (graded beds) and deducing the nature of the currents, geologists have tended to assume a particular kind of current and if some of the bed features did not fit this interpretation have argued that "turbidity currents" could not be responsible. For example, the presence of features suggesting tractional or bed-load processes (e.g. lamination, cross-lamination, paucity of matrix, etc.) within some graded beds did not seem compatible with deposition from true suspension currents, as "turbidity currents" were assumed to be, but instead seemed to infer that "traction currents" (e.g. normal ocean currents) might be

responsible. During the last decade, however, the role of tractional or bed-load processes during deposition from "turbidity currents" has been recognized, and a close similarity to "traction currents" during the depositional phases now seems established. Another major development during this period has been the realization that some process (probably dispersive pressure) other than turbulence might be responsible for maintaining the coarse grains in suspension during transport, and hence "turbidity currents" may not be true suspension currents in this sense either.

The existence of these anomalies suggests that a return to the original concept of density currents (Daly, 1936), which do not imply any particular mechanism of transport or deposition, would be more satisfactory than having to continually modify and expand the "turbidity current" concept as new insights are gained. The latter, in essence, are simply sediment-laden density underflows. Crowell et al. (1966) have already argued for a return to the more general concept of density flows in order to avoid the misconceptions and confusion surrounding the interpretation of "turbidity currents".

The realization that bottom traction has been important during the deposition of graded beds constitutes a major step in our understanding of the nature of the currents. The term "traction carpet" was coined by Dzulynski and Sanders in 1962 for the moving layer of grains which was thought to form beneath the turbidity current during deposition, and within which the lamination was thought to form. Prior to this, Hsu (1959) has used the term "fluidized sediment mass" for this layer. Two other important developments came in 1965. The first of these was the application of the "inertia-flow" principle of Bagnold (1954, 1956, 1962) to turbidity currents by both Walker (1965) and Sanders (1965). This mechanism, which involves the suspension of grains by means of the dispersive pressure created by grain collisions, was thought to operate within the basal layer as an alternative to the normal traction carpet under conditions of high bed shear (high velocity). Deposition during this phase was thought to have formed the basal structureless (i.e. unlaminated) division characteristic of most graded beds. This proposal marked the first time that any mechanism other than normal turbulence

had been seriously considered for maintaining the coarse sediment in suspension. Evidence presented in this thesis also suggests that dispersive pressure could have been of fundamental importance for transport of coarse material.

The second development was the realization that the vertical sequence of internal structures found in graded beds was almost analogous with the sequence of bed-load structures formed by waning "traction currents" in flume experiments. This analogy, which was made by both Walker (1965) and Harms and Fahnstock (1965), fitted neatly with the concept of a waning current as indicated by the graded bedding, and further suggested that there was little real distinction between "traction currents" and "turbidity currents" during the depositional phases. The basal structureless division (a) could be attributed to deposition from the upper part of the upper flow regime (bed forms include standing waves and antidunes), the division of horizontal lamination (b) to the lower part of the upper flow regime and the transition phase (plane bed with sediment movement), and the cross-laminated division (c) to the ripple phase of the lower flow regime.

Walton (1967) discussed and modified this interpretation to some extent, and more recently Allen (1969) has extended the approach in terms of flow power. Although not acknowledging the analogy, Kuenen (1966) concluded that the parallel lamination in graded beds was due to bottom traction, and he later attributed this division (b) to deposition from an "overloaded suspension current" in the upper flow regime (Kuenen, 1967, pp.223-4).

As might be expected there is not complete analogy, since the experimental work concerned the net transport of sand in subaerial currents with little or no suspended load, whereas most graded beds must be the result of net deposition from deeply submerged currents with much of their load in suspension. Perhaps the most critical problem concerns the nature of transport and deposition in the upper part of the upper flow regime, and whether the experimental water-driven flows of this type are truly analogous with sediment-driven density underflows. Results to date suggest that the two are analogous, at least to the extent that the mode of transport and the resulting deposits are

similar, even if the primary driving force is different. Little is yet known about currents of this type because of the difficulty of achieving the high velocities in laboratory flumes and of preventing reworking of the deposits by the decelerating tail of the current. Such flows are capable of transporting very high concentrations of sediment (up to 60% by weight) at considerable velocities, the sediment being maintained in a kind of suspension possibly by dispersive pressure.

A second problem concerns the absence, except in rare cases, of a division of large-scale cross-bedding within the graded bed corresponding to the dune (megaripple) phase of the lower flow regime. As explained by Allen (1969, pp.32-33), however, this is probably a function of grainsize and flow power, since in the fine sands which normally occur at this level the range of flow power over which dunes will form is very narrow. The current probably declines so rapidly through this range that there is not time for the massive redistribution of sediment necessary to form dunes. Simons *et al.* (1965, p.40) also noted that with fine sands the plane bed phase persisted down to quite low Froude numbers (0.3). One bed showing the complete sequence of internal structures, including a division of large-scale cross-bedding, is described under Group C herein, and Hubert (1966) has described some other examples. Such is the confusion regarding the nature of turbidity currents, however, that Hubert can use these examples as an argument against deposition from such currents, whereas such a sequence is to be expected under other interpretations.

b. Some Evidence from Sedimentary Features

When considering the nature of the depositing currents, it is apparent that the interpretation of the basal structureless division is of prime importance, since this was the first part of the sediment load to be deposited and hence will most closely reflect the conditions within the current during transport. The overlying laminated divisions can be roughly predicted in terms of a waning bottom current depositing partly from bed-load and partly from suspension, as discussed by Walker (1965) and Sanders (1965). The following points appear to be significant.

(i) Lack of sorting and lamination in basal division

The currents must have been capable of transporting a mixture of pebbles, sand and finer material and of depositing this material without sorting it or producing lamination except in the later stages. The presence of lamination and primary current lineation in the overlying part indicates that a fast-moving current, rather than mudflow, was involved. The little evidence available concerning upper flow regime flows (e.g. Middleton, 1965; Simons et al., 1965; Nordin, 1963) suggests that the sediment load is transported en masse, with little or no differentiation into bed load and suspended load. Nordin reported almost uniform very high sediment concentration throughout the vertical section of a natural stream with antidunes and standing waves. The resulting deposits from such flows tend to be unsorted and poorly laminated or structureless (Middleton, 1965). This feature is discussed further when the mode of emplacement of the Group C sandstones and Group D conglomerates is discussed.

(ii) Topographic control of deposition

Some beds show evidence that the initial phase of deposition was strongly influenced by bottom irregularities. Pre-depositional hollows and scour marks have been preferentially filled with coarse sediment while the adjacent high areas remained uncovered. In some cases, e.g. Bed C of Fig. 8, almost all the visible coarse-grained part of a bed was deposited in a local depression. This must mean that the depositing part of the current was in contact with the bottom but that deposition occurred only where depressions were encountered (possibly because of flow separation causing a slight local reduction in velocity). Deposition from the later phases of the current, carrying mainly silt and fine sand, appears to have been more general and blanket-like, however, and not particularly influenced by bottom topography. The Group C sandstones and Group D conglomerates also show indications of topographic control of deposition.

(iii) Transport without deposition (auto-suspension)

The presence of coarse-grained depression-fillings, as previously noted by Sanders (1965), indicates that the initial

gravel-bearing phase of the current was capable of passing over parts of the floor without depositing, except in the hollows. This is significant, since some authors (e.g. Middleton, 1966) have argued that turbidity currents carrying coarse sand and gravel could not have attained auto-suspension but must have been depositing at all stages.

The relationship between current scour marks, e.g. flute marks, and the basal coarse division also suggests that, in some cases at least, the gravel-bearing phase could pass over an area without depositing. Since the flute marks always occur beneath the basal coarse division when this is present, they were presumably formed by the earliest part of the current, prior to or during the passage of the gravel-bearing phase. Hence the presence of flute marks filled with laminated or cross-laminated fine sand, such as occur on many beds, implies that the initial part of the current bearing the coarse sediment has passed that point without depositing. The presence in some flute marks of small amounts of gravel or coarse sand overlain abruptly by the laminated fine sand of the upper divisions also indicates that the front of the current had passed without depositing.

Even longer delays in sedimentation are suggested by those beds in which remnants of a basal conglomeratic division (including isolated pebbles) are overlain directly by inter-laminated siltstone and fine sandstone of the uppermost division (e.g. Bed A of Fig. 8). These examples imply that the middle sand-bearing part of the current had also passed that point without depositing. Presumably, if the remnants of gravel were not present, there would be nothing to indicate that a gravel-bearing current had passed. Extrapolating from this, it seems possible that many currents carried coarse sediment into deeper parts of the basin without leaving any traces of it in the upcurrent areas.

(iv) Relationship to Group C sandstones and Group D conglomerates

A factor of considerable importance in considering the origin of the basal structureless division, and of the sandstones in general, is the relationship between the normal Group B sandstones and the other two groups of coarse-grained rocks, viz. the thick-bedded, usually non-graded sandstones of Group C, and the conglomerates of

Group D. The normal sandstones are transitional with both these groups, as demonstrated by the numerous intermediate types and by the examples of lateral transition from one type to the other within the same bed.

Many sandstone beds have a basal division of fine conglomerate, and there are all gradations from beds of this type to beds of conglomerate with just a thin upper division of sandstone. There are also examples of single beds which grade laterally from conglomerate to sandstone (e.g. Bed B1 of Fig. 8; Bed 3 of Fig. 10). Similarly, there are all gradations from non-conglomeratic Group B sandstone beds, in which the basal structureless division is subordinate to the overlying laminated divisions, to the sandstone beds of Group C in which the basal structureless division occupies all or most of the bed. Beds which grade laterally from one type to the other, due to variations in thickness of the basal structureless division, are common in some of the compositely-bedded sandstone zones (e.g. Plate 21A).

Thus it would appear that the same currents which deposited the normal sandstones of Group B also deposited the thick-bedded sandstones and the conglomerates. The latter types, in fact, appear to have been deposited almost wholly by the initial antidune phase of the currents, whereas the Group B sandstones were deposited partly by this phase (the basal structureless division) but mainly by the declining phases of the currents (plane bed, ripples etc.). Evidence deduced from the thick-bedded sandstones and the conglomerates concerning the nature of the depositing currents should therefore be considered before a final interpretation is given.

E. GROUP C - THICK-BEDDED SANDSTONES

1. General Features

Thick sandstone beds, ranging from one to six feet (0.3 - 2.0 m) in thickness, are common in parts of the formation and are arbitrarily designated Group C. The majority of these beds consist mainly of structureless sandstone, and many are structureless throughout. Some, however, have an upper laminated part, usually consisting mainly of parallel lamination (division b) but with the cross-laminated

division (c) also represented in many cases. Such beds are similar to the Group B sandstones, and a complete gradational series exists between the two groups, depending mainly on the thickness of the basal structureless division. The Group C sandstones also show several features, however, which are not normally seen in the thin-bedded sandstones, particularly the wavy lamination or "dish structure" in the basal division.

The sandstones are light-coloured, hard, very quartzose rocks essentially similar to those of Group B. They are predominantly medium- to fine-grained, but rare conglomeratic beds of this type also occur. Shale pellets up to 6 inches (15 cm) long occur in some beds, while others contain abundant small inarticulate brachiopod valves (apparently transported).

Beds of this type occur in many parts of the section but are particularly abundant in the coarse-grained zones in the upper part of the formation (Fig. 7). Within these zones they usually inter-tongue and inter-grade with the thinner Group B sandstones, but there are some zones which apparently consist wholly of Group C sandstone, e.g. the one forming the crest of Flagstone Knoll (Plate 19A). The zones are characterized by composite bedding, with little or no siltstone, and the contacts between sandstone beds may be marked only by thin joints or by slight changes in grain size. The intimate association of thick-bedded and thin-bedded sandstones within most of these zones, and the lateral variations from one type to the other, emphasizes the rather artificial nature of the grouping.

2. Sole Marks

Since most beds of this type occur in composite sequences lacking siltstone, sole marks tend to be rare or poorly developed. The majority of beds have either flat concordant contacts on the sandstone below, or undulating erosional contacts with a relief of up to 8 inches (Plate 21A). Thinner beds are transitional with and similar to the Group B sandstones, and may show the current marks and load pockets described for that group. Some beds have a basal part in which contorted and balled-up structures are developed, superficially



A. Composite sequence of Group C sandstones on crest of Flagstone Knoll. Note wavy lamination or dish structure accentuated by weathering; zones of flat lamination forming tops of beds; development of typical platy scree.



B. Detail of above, showing one complete bed with dish structure and parts of two others. Hammer handle rests on division of flat lamination forming top of lower bed. Contorted structure at base of upper bed can be seen above hammer.

resembling load structures (e.g. Plate 20A, B). The irregular nature of these structures, however, and the fact that they are also present within the main part of the bed, suggest that they are due to internal movements in the sand during deposition rather than to loading effects.

3. Internal Structures

Although many of the thick sandstone beds are structureless throughout, the group shows a spectrum of internal structures which include some not commonly recorded from flysch sequences. Many beds show the same sequence of internal structures as the Group B sandstones, with a basal structureless division followed by a division of horizontal lamination followed by a cross-laminated division. The upper division of laminated siltstone (d) is normally absent, however, due to truncation or non-deposition. The basal structureless division (a) is nearly always predominant, and the laminated divisions may be only intermittently developed (or preserved) along the top of the bed.

As well as these more normal features, many beds show a faint to prominent wavy or scoop-like lamination within the basal division a, and in many cases this is underlain by a thinner unit of contorted and balled-up lamination. The complete sequence, as shown by a number of beds, is as follows: (i) basal division of contorted lamination; (ii) division of wavy lamination or "dish structure"; (iii) division of plane parallel lamination, with parting lineation; (iv) division of cross-lamination and ripple mark. In one bed, a division of poorly-developed large-scale cross-bedding is preserved in the expected position between the parallel lamination and the upper division of ripple cross-lamination. The structure is only intermittently developed along the bed, however, and no other examples have been seen.

a. Contorted Lamination

A division of contorted and balled-up structures is developed in the basal part of many beds, and normally grades upwards into dish structure (e.g. Plate 20A). The contorted laminae usually consist of slightly darker, more micaceous sandstone. The structures are concave upwards, and range in size from a few inches in diameter to 12 inches



A. Single Group C sandstone bed showing contorted structure at base merging upwards into dish structure, followed erosively by lighter-coloured and finer-grained division of flat lamination.



B. Same bed as above, 10 feet to south. Note basal contorted structure, apparent absence of dish structure, thin division of flat lamination merging into division of poorly-developed cross-bedding.

contact in many cases is undulating rather than planar. On some beds it is only intermittently developed and appears as broad shallow troughs a few inches deep. Similarly, the relative proportion of the bed occupied by this division may vary considerably, particularly in those beds which show rapid lateral variations in thickness (e.g. Plate 21A).

d. Ripple Mark

The tops of many Group C beds have a thin division of ripple mark and cross-lamination, generally only an inch or so thick. The ripples are generally rather irregular in form, consisting of oval or sub-circular troughs separated by rounded crests. Linear ripples have not been observed on these beds.

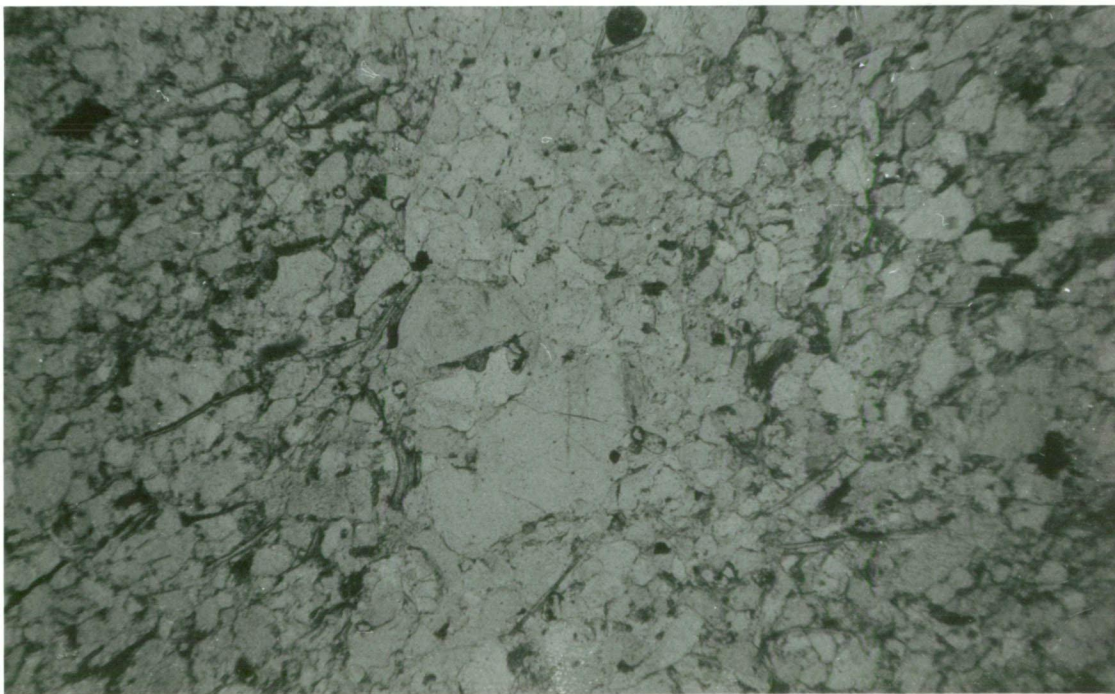
e. Large-scale Cross-bedding

In one bed the division of parallel lamination is overlain by, and apparently laterally gradational with, a division of poorly-developed large-scale cross-bedding (Plate 20B). This in turn is overlain by ripple cross-lamination. The cross-bedded division is up to 10 inches (25 cm) thick, but is not continuous along the bed. Two interfering sets are present in places, the upper one eroding completely through the lower one (Plate 20B), but elsewhere there is only a single set about 6 inches thick. The foresets have a maximum inclination of only 18° , and are tangential to the base where they appear to grade into the plane parallel laminae.

The cross-bedding appears to have been formed by the filling of shallow-troughs by accretion of sub-parallel laminae rather than by the forward migration of avalanche faces as in normal dune cross-bedding. Imbrie and Buchanan (1965) have described a somewhat similar kind of cross-bedding ("accretion deposits") which they considered to be intermediate between sheet deposits (plane parallel laminae) and avalanche deposits (normal foreset cross-bedding). Jopling (1965) has shown experimentally that cross-laminae become progressively flatter and more tangential to the base as velocity is increased.



A. Composite sequence of mainly Group C sandstone beds. Note marked wedging of middle bed due to thinning of basal structureless division; also thin dark contact line between lower two beds, contorted structure at base of upper bed.



B. Photomicrograph of elutriation column in Group C sandstone. Note absence of mica, coarser grainsize, upturning of adjacent mica flakes. Dark rounded grain is tourmaline. Ordinary light. Height of field is about 2 mm.

f. Elutriation Columns

Some of the Group C beds, particularly those in which dish structure is well developed, show numerous, irregular, sub-vertical "veins" of light-coloured mica-free sandstone. These structures are usually only a few mm wide, but may persist laterally for over 50 cm before disappearing. They are best seen on stratification surfaces, where they tend to project slightly and resemble worm tubes. They occur mainly in the upper parts of beds showing dish structure, but do not appear to be developed within the parallel-laminated division when this is present. They persist in depth for at least 10 cm (4 inches) in some cases, but their precise relationship to the dish structure has not been determined.

As seen in thin section (Plate 21B) the "veins" consist of pure quartz sandstone lacking micaceous material and fine grains. In some cases they contain distinctly coarser grains than the surrounding sand (Plate 21B). Faint lamination in the adjacent sand, as indicated by oriented mica flakes, may be dragged upwards against the sides of the structures, suggesting upward movement of material within the veins. The structures are probably equivalent to the "elutriation columns" mentioned by Wentworth (1967), and are attributed to localized upward movement of water through the bed during deposition, causing washing-out of the fines and some upward transport of coarse grains.

4. Graded Bedding

Although the majority of Group C beds are more or less uniform in grainsize throughout, there are also a number which show graded bedding. The most spectacular examples are those beds which have a conglomeratic base followed by dish structure, but such beds are rare. Most of those beds which have an upper division of parallel lamination show graded bedding, since this division is nearly always finer-grained than the underlying sand. The uppermost division of ripple cross-lamination, where present, is usually finer still. An abrupt break in grainsize, accompanied by an improvement in sorting, usually occurs at the beginning of the parallel-laminated division (Plate 20). The improvement in sorting is partly due to the loss of much of the

micaceous material, and this is apparently what causes this division to be lighter-coloured than the underlying sand. This same break is also typical of the Group B sandstones.

5. Lateral Variations

Although outcrops are mostly too restricted to allow tracing of individual beds for more than a few tens of feet laterally, it is apparent that some beds at least show considerable variations in thickness and structure. Some of the thinner beds can be seen to wedge out almost completely within a few feet (e.g. Plate 21A), and there are many beds which double or treble in thickness over similar distances. Both erosional and depositional lensing appear to be responsible, and in places there is a tendency towards an imbricate arrangement such that a bed reaches maximum thickness where the preceding bed is thinnest. There are many beds, however, which maintain a more or less uniform thickness across the outcrop width.

Lateral variations in internal structures are also common. The intermittent occurrence of a division of parallel lamination along the top of some beds has already been mentioned. Those beds which show marked changes in thickness may be represented only by this division of parallel lamination where they are thinnest (e.g. Plate 21A). The thinner parts of such beds could be classified as Group B. Lateral changes from dish structure to structureless sand are typical, and the form of this structure, i.e. whether scoop-like or sub-horizontal, is also variable.

6. Composition and Texture

The Group C sandstones are similar in composition to those of Group B, i.e. they consist predominantly of quartzose grains with a small to moderate amount of mica (mainly muscovite) and minor amounts of non-quartzose rock fragments (schist, slate, dolomite), carbonate, and accessory minerals (particularly tourmaline). Modal analyses of two samples from the crest of Flagstone Knoll are given in Table 1 (C1, C2). Both are from beds showing dish structure, C2 being from the upper part in which elutriation columns are developed. They differ from the typical Group B sandstones mainly in their higher mica content.

This is perhaps to be expected, since most of the Group B samples were from the parallel-laminated division, from which much of the mica was apparently removed during deposition.

Texturally, the Group C sandstones appear to encompass the same wide range as the Group B beds. Although the two samples for which grainsize data were obtained (C1, C2 in Fig. 9) are both fine-grained and well-sorted, there are also coarse-grained and conglomeratic beds which must be poorly sorted and texturally immature. A marked improvement in sorting occurs in the parallel-laminated division when this is present.

7. Previous Descriptions of Similar Sandstones

Unusual thick-bedded sandstones, some showing dish structure, have been described from flysch sequences by Wentworth (1967, abstract), and Stauffer (1967). Crook (1961) recorded graded beds showing "discontinuous curved lamination". The beds described by Wentworth were generally about one metre thick, and typically showed a coarse-grained basal structureless division passing upwards through flat lamination, dish structure and overlying flat lamination to a top of fine sandstone with convolute lamination. Some beds showed dish structure as a central division within an otherwise structureless bed.

Stauffer (1967) described and illustrated similar sandstones and coined the term "grain-flow deposits" for them. The sandstones were thick-bedded, mostly non-graded, massive except for dish structure or diffuse flat lamination, and contained large shale clasts in some cases. Some beds showed unusual sole structures such as frondescant marks, drag and slide marks, load structures, and ropy sole marks. Stauffer rejected an origin from "classical" turbidity currents because of the "very large outsize clasts and the lack of grading and other expected structures". He proposed instead a mass-flow mechanism dependent on the dispersive pressure of grain collisions (Bagnold, 1954, 1956). Such flows were thought to form independently of any turbidity current and to be triggered by slides on a steep slope. Evidence from the present rocks suggests, however, that the graded and non-graded sandstones were deposited by the same currents, and that the mechanism

proposed by Stauffer was an integral part of the normal density currents rather than a new and different kind of flow. It is recommended that the term "grain-flow deposits" be dropped because it involves the assumption of a mechanism which is poorly known and which, in any case, was probably part of a more general kind of flow which also produced normal graded beds.

8. Origin of Dish Structure

Wentworth (1967) attributed the dish structure to antidune flow as follows: "Antidunes may produce simple dish structure by alternate scour, and be breaking and deposition in the troughs during aggrading suspension flow in a turbidity current that has declined from dispersion flow (after Bagnold). ... Some characteristics, possibly including clay distribution, may result from sporadic water expulsion, related sediment flowage, and general de-watering of the bed."

As noted by Wentworth, the antidune "cross-bedding" produced experimentally by Middleton (1965) is very similar in size and shape to dish structure. The "cross-bedding" consisted essentially of a series of slightly concave lenses a few cm deep and 30-150 cm (1-5 feet) wide. Characteristics of the structure were the faintness of the lamination, the low angle of the cross-bedding, and the complete absence of any foreset (angle-of-repose) cross-laminae. The production of the lenses appeared to depend on the breaking of the surface (water) wave, causing it to be temporarily displaced upstream from the sediment wave and resulting in a separation zone just upstream of the latter. A lens of sediment was deposited in this zone and was partly preserved when the former sediment wave (antidune) was eroded. The preserved lens was somewhat shorter than the wavelength of the original antidunes.

No other mechanism for producing this peculiar structure is known, and the supposition of antidune flow agrees with other hydrodynamic reconstructions of turbidity currents (e.g. Walker, 1967, 1965) and with the evidence from other sedimentary features.

Wentworth also inferred that the structureless division which occurs below (and sometimes above) the division of dish structure in some beds represents deposition from "dispersion flow" (presumably flow

in which the sediment load is suspended by means of Bagnold's dispersive pressure) before the current declined to antidune flow. The probable importance of dispersive pressure in upper flow regime flows will be discussed with respect to the Group D conglomerates, but the assumption that it represents a separate phase of flow above antidunes may not be justified. The poorly-defined nature of dish structure and its absence from many beds suggest that its preservation is somewhat fortuitous, and that the more normal deposit from antidune flows would be structureless sand. Flume experiments indicate that flow in the upper flow regime is likely to be spatially irregular, with chutes and pools separating trains of antidunes (e.g. Simons et al., 1965), and hence any structures formed as the antidunes break are likely to be irregularly distributed along and within the bed.

F. GROUP D - CONGLOMERATES

Quartzose conglomerates, mostly of pebble grade but up to boulder grade in places, occur sporadically throughout the formation and are predominant in a few places. The conglomerates are usually interbedded with sandstones, and are similar in colour and composition to the sandstones. All gradations exist from beds of conglomerate with a thin sandy top to beds of sandstone with a thin basal division of conglomerate, and because of this overlap there is little point in rigidly defining the group. The present discussion concerns those beds composed mainly of conglomerate and which would be called conglomerates in the field.

1. Distribution

Single beds of conglomerate, ranging in thickness from a few inches to 3 feet (90 cm) or so, are interbedded with sandstones at many localities, particularly within the coarse-grained zones or lenses. Conglomerate beds isolated within siltstone are rare, however. In the lower half of the formation there are several zones in which conglomerate is the predominant rock type. One of these is exposed at the foot of the range two miles north of Flagstone Knoll (Fig. 5), and another on a small ridge above the Boyes River adjacent to Flagstone Knoll.

Neither zone appears to be continuous laterally, but confirmation of this is difficult because of lack of outcrop. A smaller zone, in which conglomerate beds are well exposed over a short interval, is exposed on the south flank of Flagstone Knoll near the head of Singing Creek. Here again, the outcrops suggest that neither the zone nor the individual beds are laterally continuous. Most of the following discussion is based on these three areas, since the features shown by these rocks are typical of the conglomerates in other parts of the formation.

2. Composition

The conglomerates consist predominantly of quartzose clasts in an abundant sandy matrix (Plate 18D). Micaceous silty material is present within the matrix, but seldom exceeds 10% of the total rock. Modal analyses of three of the fine conglomerates, and of the sandstone part of a conglomerate bed, are given in Table 1. The proportion of clasts varies greatly, and there are all gradations from conglomerate with 70-80% clasts to sandstone with only scattered pebbles. The clasts in most beds do not exceed pebble size, and for the most part are less than one inch (2.5 cm) in diameter. Only in the coarse conglomerates north of Flagstone Knoll is there a significant proportion of cobbles and boulders (Plate 22A).

At least 90% of the clasts are composed of either quartzite, quartz-schist or vein quartz, and clasts larger than pebble size are almost exclusively of these types. Other rock types are usually only apparent in thin section, and include quartz-muscovite schist, quartz-sericite schist, chloritic schist, dark phyllite or slate, chert, and rare dolomite. Intraformational fragments of dark micaceous siltstone and light-coloured fine sandstone are present in most beds, and are very abundant in some. Many of these are elongated and deformed (e.g. Plate 24B). The sandy matrix is also predominantly quartzose, and is similar to the Group B sandstones in composition.

3. Texture

Textural studies of the cobble-grade and coarser conglomerates have not been attempted, since these rocks cannot be disaggregated or effectively studied in section. Visual examination indicates that a

complete range of grainsizes from boulders to fine mica is present, but the precise nature of the grainsize distribution is not known. The fine conglomerates can be examined in thin section. Examination of hand specimens of these strongly suggests that they are bimodal, the pebbles and granules forming one mode and the medium to fine sandy matrix the other, and thin sections confirm this (e.g. Plate 18D). The micaceous material probably forms a third mode.

The grainsize distributions of four thin sections from the fine conglomerates have been determined from modal analysis, by the method described earlier for the Group B sandstones, and are shown in Fig. 9. Three of these (D1, D2, D4) are from the locality on the south slope of Flagstone Knoll (the beds are shown in Fig. 10), and one from the ridge above the Boyes River (D3). The curves clearly show the wide range of grainsizes present within the conglomerates, and the bimodal distribution of the quartzose grains. Another feature of the fine conglomerates which is reflected in the graphs is the relatively good sorting of the coarse mode, the bulk of which consists of small pebbles and granules. Specimen D4 is from the upper sandy part of a conglomerate bed shown in Plate 26A and Fig. 10, and is a well-sorted fine sandstone similar in composition and texture to the typical Group B sandstones.

The cutoff between the gravel and sand modes appears to vary somewhat from bed to bed, although this may be a function of the poor representation given by the thin sections. The main cutoff, however, is between one and two mm (-1 to -2 phi), and there is a paucity of coarse and very coarse sand in the conglomerates measured. However, these sizes predominate in the lower parts of some graded sandstone beds (e.g. B2(a), B2(b) of Fig. 9), and it is doubtful if there is actually an overall paucity of these grains in the formation. Much more work is required to clarify these points.

Roundness measurements have not been done. The pebbles and large clasts range from sub-angular to well-rounded (Plate 18D), the majority being sub-rounded to rounded. Sphericity varies from low to high, since the more schistose fragments tend to be platy or tabular while many of the quartzite clasts show high sphericity. The roundness of the sand-grade material varies greatly, as it does in the Group B



A. Graded conglomerate beds two miles north of Flagstone Knoll. Note cobbles and boulders of quartzite, "floating" clast in upper bed.



B. Upper surface of graded conglomerate bed showing projecting quartzite boulder near hammer. Same locality as above.

sandstones. Most of the grains tend to be sub-angular to sub-rounded, but angular and very well rounded grains are also common.

4. Coarse Conglomerates North of Flagstone Knoll

The coarsest conglomerates crop out on a low saddle (7648N, 4237E) about two miles north of Flagstone Knoll, within a few hundred feet of the base of the formation. Access to this area is difficult, and the outcrops were visited only briefly by the author. Some 10-20 conglomerate beds, each 4-6 feet (1-1.5 m) thick, are exposed over a stratigraphic interval of about 150 feet, but the interbedded rocks are not exposed. Individual beds appear to be tabular, but the majority cannot be traced along strike for more than a few tens of feet because of lack of outcrop.

Nearly every bed is distinctly graded from cobble or boulder conglomerate at the base to pebbly coarse sand at the top (Plate 22A, B). Boulders of quartzite up to at least 18 inches (35 cm) in diameter occur in the lower parts of some beds. The grading is seldom regular, in that boulders may occur at any level in the bed, and in some cases single boulders project through the upper surface of the bed (Plate 22B). However, there is usually a pronounced decrease in the proportion of large clasts upwards.

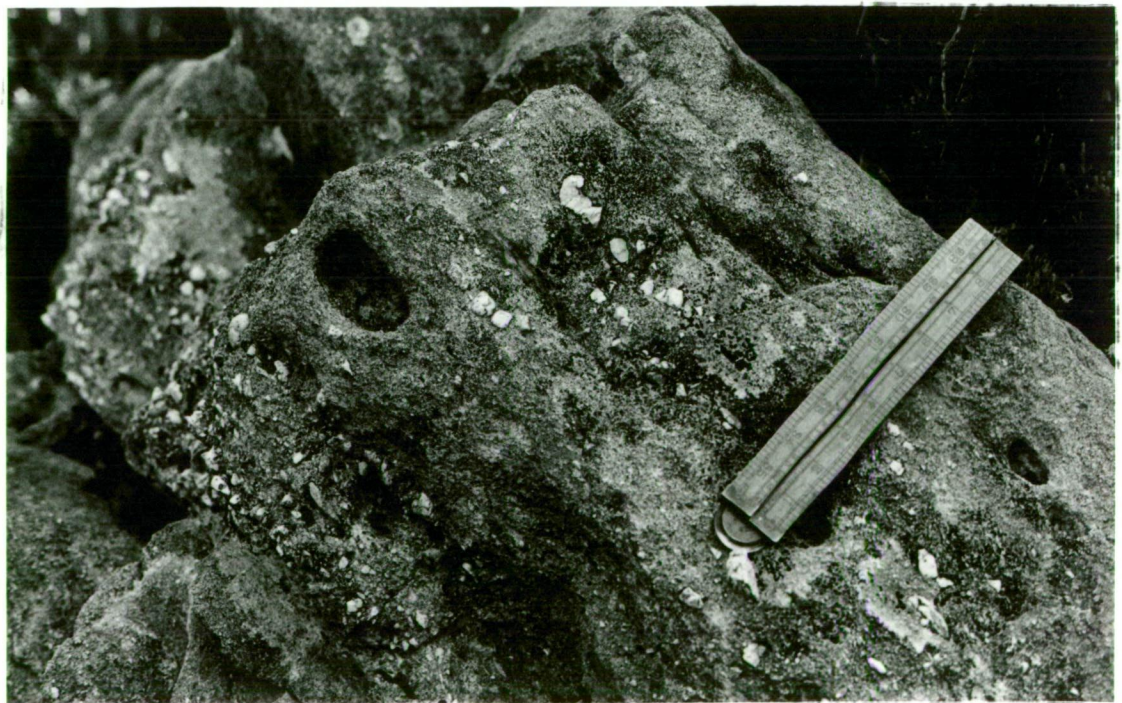
The exposed soles of the beds are very irregular because of the megaclasts present, but the presence or otherwise of sole marks (channels, grooves etc.) could not be determined because of lack of exposure. The exposed upper surfaces, on the other hand, are relatively smooth and planar (Plate 22B). Internal structures, apart from the graded bedding, are seldom apparent, although a few beds show a faint tendency towards horizontal stratification in the sandy upper part.

5. Ridge Above Boyes River

Conglomeratic rocks are exposed along the crest of a small spur (7605N, 4234E) just southwest of Flagstone Knoll near the headwaters of the Boyes River (Fig. 5). These beds are finer grained and less regularly bedded than those described above. They occur within a stratigraphic interval of about 150 feet (Fig. 7), above a sequence consisting mainly of siltstones. The zone extends along strike for



A. Graded fine conglomerate bed, ridge above Boyes River.



B. Poorly-graded fine conglomerate bed with "floating" pebbles. Note holes left by intraformational siltstone clasts. Ridge above Boyes River.

about half a mile, but individual beds can seldom be traced for more than a few tens of feet because of lack of outcrop. The conglomerates are interbedded with sandstones of both Group B and Group C types, but these are poorly exposed.

The ledge-like outcrops project up to ten feet above groundlevel, and may consist of single beds (Plate 23A) up to four feet (1.2 m) thick, or composite units containing several beds (Plate 24A). Clasts do not exceed pebble size, the largest observed being 2.8 inches (7 cm). Fragments of intraformational siltstone and fine sandstone are common in some beds, and may occur at any level within the bed (e.g. Plate 23B). Many of the beds are distinctly graded from pebble conglomerate at the base to coarse sand at the top (Plate 23A), but poorly graded (Plate 23B) and apparently non-graded beds also occur. In some cases the grading is evident only as a fairly abrupt change to sandstone near the top of the bed.

Most of the beds are non-laminated or show only a faint tendency towards parallel lamination in the upper part. In some cases, however, there are fairly distinct alternations of sandy and pebbly bands, up to a few inches thick, within a bed (Plate 24A), and these form a crude parallel lamination. Cross-bedding has not been observed. Tops and bottoms of beds are mostly flat across the outcrop width, but shallow undulations, possibly representing erosional grooves, occur on the soles of a few beds.

6. Head of Singing Creek

A zone of interbedded sandstone and conglomerate is intermittently exposed on the southeast flank of Flagstone Knoll near the head of Singing Creek. The zone is about 12 feet thick (Plate 24B) and persists along strike for at least 200 feet. The base is exposed only at one point, where it consists of a conglomerate bed with a sharp (erosional ?) contact on the underlying sequence of fossiliferous siltstones. The middle part of the zone consists mainly of sandstone beds of both Group B and Group C type, some of them with bands of fine conglomerate. These beds range in thickness from an inch or so up to about 18 inches (45 cm). A few thin siltstone bands are also present,



A. Composite sequence of conglomerate and sandstone, ridge above Boyes River. Note crude lamination in upper part.



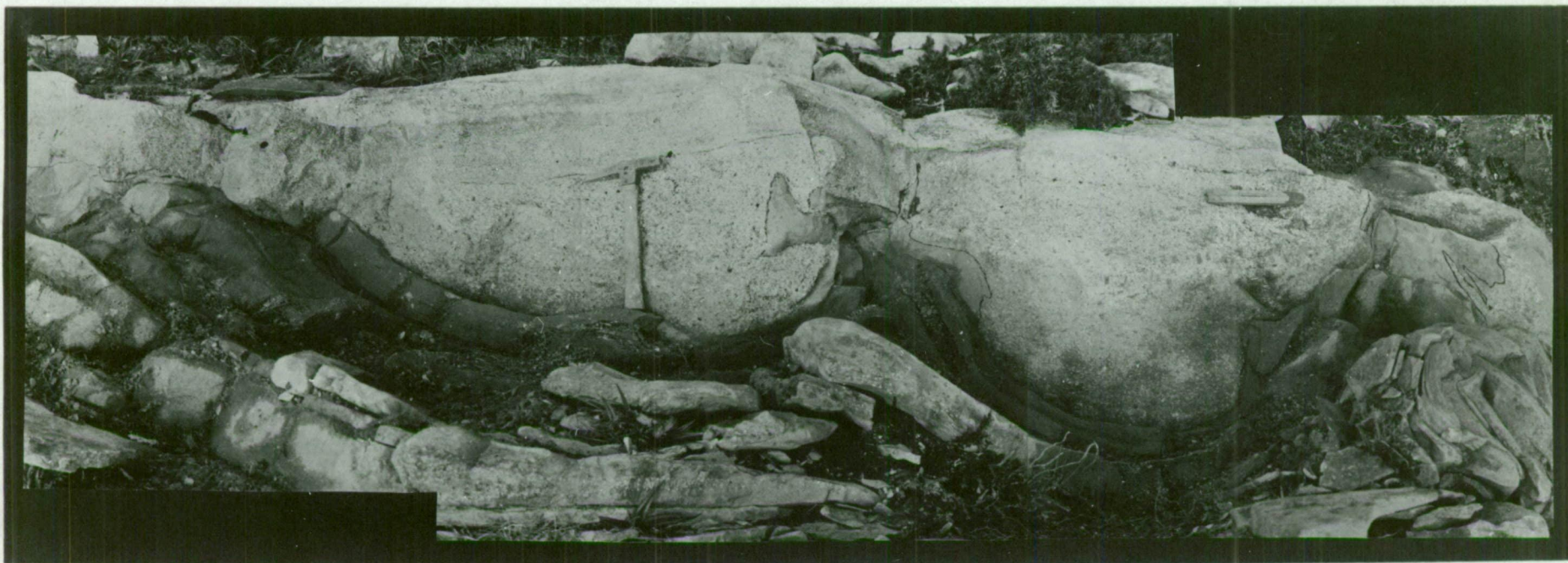
B. Conglomerate-sandstone zone near head of Singing Creek. Note shale pellets in basal bed, zone of folded sandstones beneath upper conglomerate.

and siltstone fragments are abundant in some of the sandstone beds (Plate 24B). The upper part of the zone is conglomeratic, and consists of several interfering beds of fine pebble conglomerate showing marked variations in thickness (Fig. 10). The zone is overlain by a sequence consisting mainly of siltstones with some thin sandstones and slump sheets.

The lower conglomerate (Plate 24B) is about 15 inches (38 cm) thick, but is exposed for only a few feet laterally. It consists of quartzose pebbles up to one inch long in an abundant sandy matrix. Numerous elongated and contorted siltstone fragments, up to 15 inches long, occur within the central part of this bed, and smaller fragments of fine sandstone are also present. The upper few inches of the bed are finer-grained and better sorted than the remainder, and show a faint parallel lamination. The exposed base of the bed is flat and smooth.

The conglomerate beds at the top of the zone are well exposed in several outcrops (Plates 25, 26), but the limited size of the exposures makes it difficult to determine the precise shape of individual units. A scale section of the best exposed area, where three conglomerate beds are partly exposed over a strike length of about 43 feet (13 m), is shown in Fig. 10. Plates 25 and 26A show parts of this section, and Plate 26B shows the same horizon further north. Modal analyses and grainsize distributions of three samples from these beds are given in Table 1 and Fig. 9 (see Fig. 10 for sample locations and numbers).

Although at least four conglomerate beds are represented within this section, only one or two of these reach their maximum thickness in the same area. As shown by Fig. 10, there is a marked tendency towards an imbricate arrangement, with successive beds reaching maximum thickness where the preceding bed lenses out or becomes markedly thinner. Most of the thickness variations are due to depositional lensing, but erosion of beds is also apparent in places (Plate 26B). Undulating erosion surfaces, with a relief of at least 8 inches (20 cm), occur beneath some beds in the areas where the underlying bed is thickest. The imbricate arrangement suggests that the gravel was deposited as irregular mounds or lenses rather than as flat sheets and



Upper part of zone shown in Plate 24B. Conglomerate bed with large deformational sole structures affecting underlying sandstones. Note more intense deformation in right-hand anticline, ends of disrupted sandstone beds at left. Current from left to right. See Fig. 10.

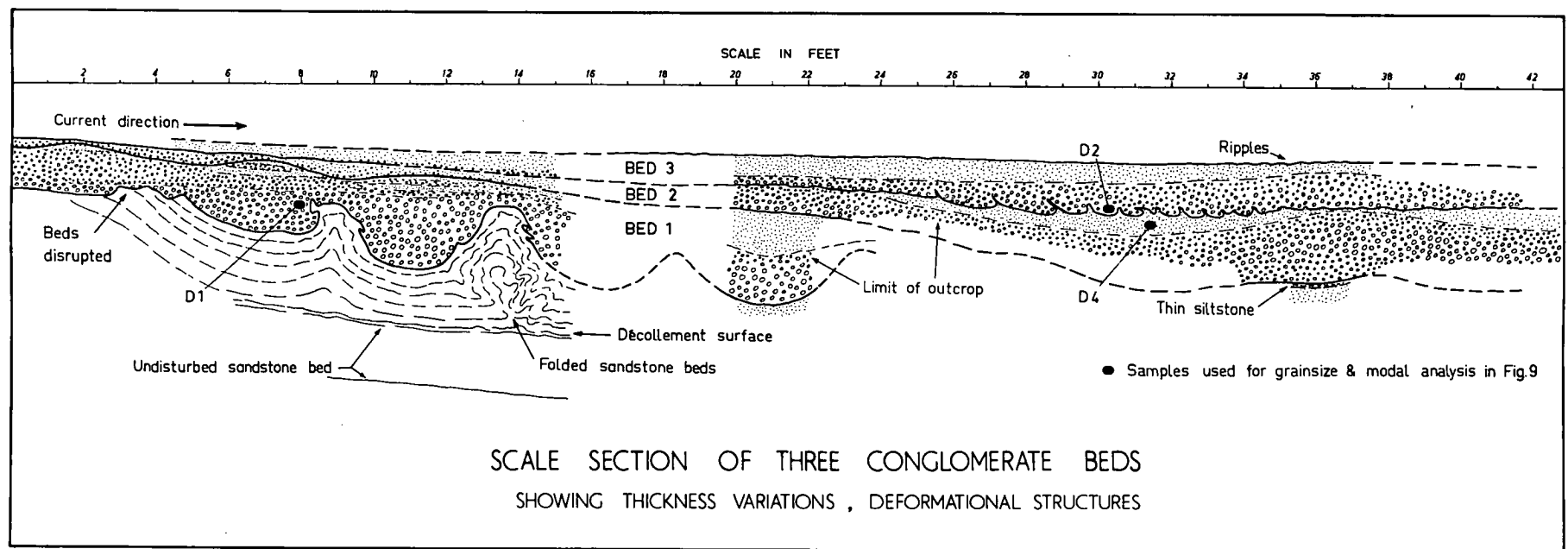


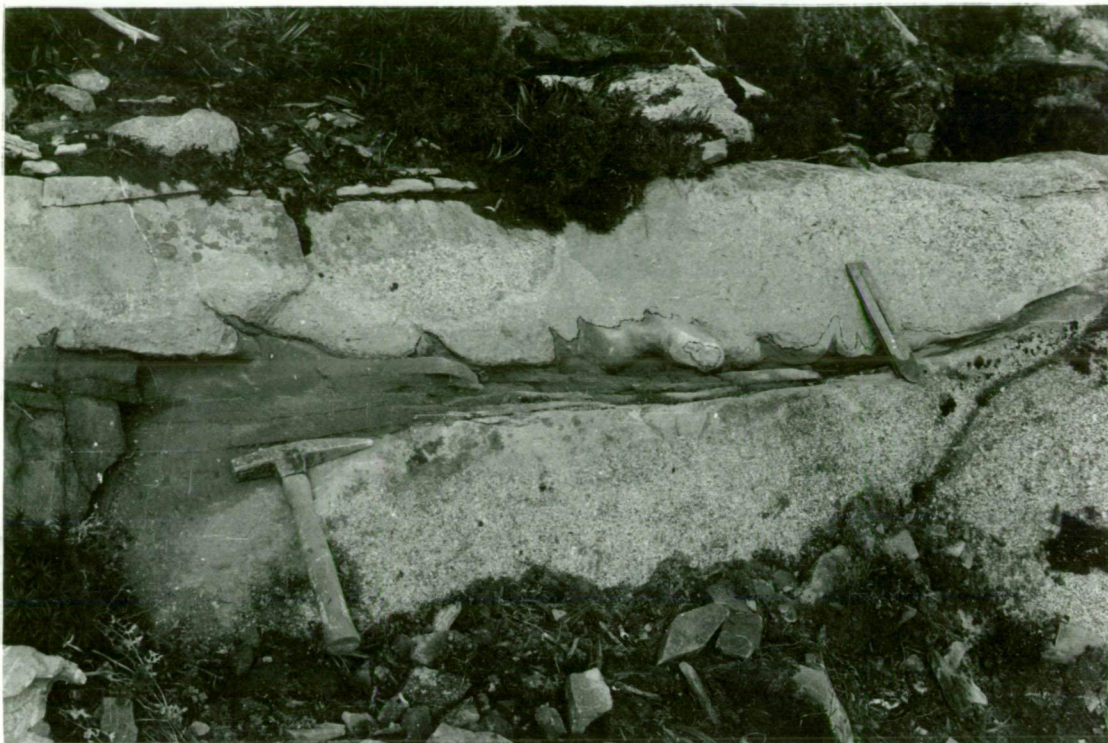
FIGURE 10

that each current tended to erode the top of the previous lens and then deposit its load in the low area in front of the lens. This topographic control of deposition suggests that the bulk of the sediment load was being carried as a dense suspension in contact with the bottom.

Most of these beds are distinctly graded from conglomerate at the base to sandstone at the top. The grading is nearly always of the delayed or interrupted type, however, with sharp but usually gradational contacts between the gravel and sand horizons. In some cases, as in the bed shown in Plate 25, there is a decrease in grainsize of the gravel component upwards, but again the grading occurs as a series of steps separated by fairly sharp, sub-horizontal contacts. In most beds the gravel component is fairly uniform throughout, and most of the grading occurs at the gravel-sand boundary. The sandstone component ranges from coarse to fine-grained, and in some cases there is an upward decrease in grainsize. Sample D⁴ of Fig. 9 is from the sandstone part of Bed 2 in Fig. 10 (see also Plate 26A) and is fine-grained and well sorted like the typical Group B sandstones.

Within the thicker parts of some of the conglomerate beds there is a tendency towards a complementary lens-like arrangement of the gravel and sand components. The coarse material forms broad mounds or lenses, while the sandstone occurs mainly in the hollows between these and tends to smooth off the upper surface of the bed (e.g. Beds 2, 3 in Fig. 10). Such beds may consist wholly of conglomerate in one place and wholly of sandstone a few tens of feet away, and this, combined with the marked thickness variations, makes correlation between outcrops extremely difficult.

Many of the beds show internal stratification of some kind, particularly in the upper sandy parts. The stratification is most evident as alternating sandy and conglomeratic bands (Plate 25), but a faint to prominent flat lamination is also present in the pure sandstone parts of many beds (Plate 26A). The conglomeratic division also shows a faint stratification in some cases, due to slight variations in the amount of sandy matrix or to concentrations of pebbles or shale pellets. A faint imbrication of tabular clasts and of pellets of shale and sandstone occurs in a few places in some beds. The upper surfaces of a few



A. Beds 2 and 3 of Fig. 10. Note graded bedding, sole structures, complementary lens-like arrangement of sand and gravel within beds.



B. Approximately same horizon as above, about 100 feet to north. Note strongly erosional contact at base of upper bed, crude lamination in conglomerate.

beds show non-linear current ripple marks, but large-scale cross-bedding has not been observed.

Deformational sole structures

Two of the beds shown in Fig. 10 show deformational sole structures. On Bed 3 (see Plate 26A) these consist of a series of small lobes 3-10 inches (8-25 cm) wide and up to 5 inches deep separated by pointed flame structures formed by the underlying sand. The lobes tend to be flat-bottomed and elongated perpendicular to the outcrop surface (and perpendicular to the current direction), but some at least are three-dimensional with closed ends. The flame structures are mostly either vertical or steeply inclined to the left (up-current), possibly suggesting a genetic relationship to the depositing current rather than simple load deformation.

Bed 1 shows a series of much larger basin-like structures which affect the underlying thin-bedded sandstones to a depth of nearly two feet (Plates 24B, 25; Fig. 10). The sandstones are folded concordantly with the basin-like structures to form broad synclines separated by sharp anticlines. The basins thus resemble large load structures, and are up to 30 inches (75 cm) deep with a wavelength of about 4.5 feet (1.35 m). They are elongated in a northeast direction, which is perpendicular to the current direction as indicated by ripple marks and imbrication. The anticlines do not persist into the upper part of the bed, but are overlain by a division of undisturbed horizontal lamination. This suggests that the structures are syn-depositional and that their formation was completed before the top of the bed was deposited.

Deformation within the underlying sandstones is most intense in the anticlines, and the amount of deformation apparently increases to the right since the right-hand anticline is strongly compressed and overturned (Plate 25). The deformation does not die out gradually downwards, but stops abruptly at a bedding surface above a thick sandstone bed (Plate 24B), indicating decollement-type movement. This is also indicated by the fact that the sandstone beds, including those immediately beneath the conglomerate, show little or no thinning in the anticlines. Thus the increase in length of the beds necessary to produce the

anticlines must have involved bedding-plane slip and probably disruption (pull-apart) of the beds further up-current. The disrupted ends of some of the sandstone beds are exposed beneath the conglomerate just to the left of the first basin (Plate 25). The step-like nature of the contact in this area, with the uppermost sandstone beds displaced further to the right, is also to be expected, since these beds must have been dragged further to form the higher parts of the anticlines.

The folding is interpreted as having been caused by the rapid deposition of gravel from a current moving from left to right (NW to SE). The underlying beds were deformed into a series of anticlines, and the deformation associated with the development of these folds, combined with the forward push exerted by the gravel before it came to rest, caused the disruption, sliding and piling-up of the beds in the down-current direction. The anticlines may have been initially developed because of the shearing stress imposed on the bottom by the over-riding current, and could thus be analogous to antidunes. The scale of the structures and their tendency to point or "break" up-current support this analogy. If they are antidunes, then substitution in the equation relating velocity and antidune wavelength given by Middleton (1965) gives a current velocity of 4.8 feet per second or 1.45 metres per second, which agrees well with values given by Nordin (1963) for natural flows, and with those calculated by Walker (1967) from bed features in a turbidite.

7. Mode of Emplacement of the Conglomerates

a. Background

Similar conglomerates occur within many flysch deposits throughout the world. They may be graded or non-graded, usually contain a modest amount of fine material, and are commonly associated with slump deposits. The name "fluxoturbidite" was coined for such beds by Dzulynski et al. (1959, p.1114) "because characteristics of deposition from turbidity currents appear to be mixed with evidence for sliding". Walker (1967b) has summarized the features of "fluxoturbidites" and demonstrated that they are essentially those of proximal turbidites. He recommends, therefore, that the term be dropped in favour of a

descriptive term such as conglomerate, pebbly mudstone, etc., a view shared by the present author. Excellent descriptions of well-exposed conglomeratic flysch sequences are given by Scott (1966) and Unrug (1963).

Various mechanisms have been tentatively proposed for the emplacement of the conglomerates, including turbidity currents, slides, and normal ocean currents, the most generally accepted being some process intermediate between sliding and true suspension flow. Thus Dzulynski et al. (1959, p.1114) postulated "a turbidity current in which most of the sand and gravel moves in a watery slide along the base. The current is too poor in clay to raise this load in suspension, and the slope is too steep for the load to come to rest until it has spread out in a layer". Unrug (1963) and Marschalko (1964) also proposed "watery slides". Rizzini and Passega (1965) thought the deposits were intermediate between those of storm waves and turbidity currents, and called them "undaturbidites".

Scott (1966) suggested that the graded conglomerates could be due to high-velocity turbidity currents, while non-graded units could represent "either part of a continuous series of gravity-induced flow conditions, ..., or they consist of material reworked or distributed by normal marine currents" (ibid, p.100). The supposition of ocean currents was prompted because of the paucity of fine matrix and because there was some disparity between the deduced palaeoslope and palaeo-current directions.

The lack of agreement regarding the mechanism stems mainly from the association of features suggesting suspension transport (graded bedding, generally poor sorting etc.) with features suggesting sliding (included blocks of shale and sandstone, deformation of underlying beds, presence of large clasts etc.) and features suggesting traction (rare cross-bedding and imbrication, paucity of mud). It seems possible that all these features can be reconciled with one kind of flow in which most or all of the material is carried in a dense suspension because of high velocity. Such flows occur in ephemeral streams, where they are typified by antidune development, but their occurrence in marine environments can only be surmised.

b. Sedimentary Features

Consideration of the following sedimentary features is relevant to the deduction of the mode of emplacement of the conglomerates.

(i) Association with Group B sandstones

The fact that many of the Group B sandstone beds contain pebbles or layers of conglomeratic material indicates that the currents which deposited the sandstones were also capable of transporting gravel. Since there is a completely gradational series from sandstone beds with a thin basal division of conglomerate to conglomerate beds with a thin upper division of sandstone, and since some beds show lateral transitions of this kind, it seems most likely that the same currents were responsible for both types. Thus the deductions regarding the deposition of the sandstones are also applicable to the conglomerates.

(ii) Poor sorting

The poor sorting of the conglomerates, with intermixed pebbles, sand and silt, indicates that the sorting action which accompanies deposition in the tractional regimes below the antidune phase (i.e. plane bed, dunes, ripples) probably did not occur. This lack of sorting is considered to be characteristic of antidune flow (Middleton, 1965; Harms and Fahnstock, 1965; Simons et al., 1965), and is attributed to the fact that the sediment load is transported as a single mass with little or no segregation into bed load and suspended load, and little opportunity or time for sorting processes to operate. The feature is difficult to reconcile with any known kind of normal ocean current.

Closely allied with the lack of sorting is the lack of well-defined lamination or stratification within the conglomerates, except in the upper sandy intervals. Although Middleton (1965) has described faint lens-like antidune bedding in experimental fine sands, he considers that such structures would "not be visible in a coarser sediment, or in the type of unsorted sediment found in some turbidites" (ibid, p.923).

(iii) Upper sandy division with flat lamination

Many of the conglomerate beds show a fairly prominent flat lamination in the upper sandy part, accompanied by an improvement in sorting. This would be expected from a waning flow which passes from

the antidune phase to the plane bed phase of the upper flow regime as the velocity decreases. The ripple marks and cross-lamination which occur at the tops of some beds complete the expected sequence, as in the Group B sandstones, and indicate lower flow regime conditions. Dune-phase cross-bedding has not been observed.

(iv) Distribution of sand and gravel within beds

The rather irregular lens-like distribution of the gravel in some beds, with the intervening hollows filled with sand, is also suggestive of deposition from the upper flow regime (antidunes or chutes and pools). Middleton, (1965, p.924) observed that with chutes and pools in the flume "part of the bed of the flume was swept almost clear of sand (the chute) and in other parts of the flume the sand accumulated and formed large antidunes". The resultant deposit consisted of long mounds of sand within which the faint secondary lamination was developed. Further deposition, which occurred in the plane-bed phase, tended to fill in the hollows with parallel-laminated sand. The irregular nature of antidune flow, with trains of waves developing and breaking or migrating, is well known and suggests that the sediment deposited from such flows would tend to have a lens-like arrangement.

(v) Deformational structures

The lateral deformation of the thin-bedded sandstones beneath Bed 1 of Fig. 10 indicates that deposition was rapid and sudden and that the gravel retained considerable lateral momentum. Such features must imply a dense, high-velocity flow moving close to the bottom. The large basin-like structures developed on this bed indicate non-uniform deposition of gravel in a wave-like pattern due to development of anticlines in the underlying bed material. Although antidune structures have not yet been produced experimentally in semi-cohesive bed materials, the scale of the structures and their general form suggests they could represent "frozen" antidunes.

(vi) Topographic control of deposition

This feature has already been discussed with respect to the Group B sandstones, where the presence of coarse-grained depression-fillings on the soles of beds indicates that the currents were

bottom-hugging and that deposition was influenced by the presence of low areas. The tendency towards a downcurrent imbrication of successive conglomerate beds, such as shown in Fig. 10, also suggests that deposition tended to occur preferentially in depressions, while adjacent high areas were partly eroded off.

c. The Coarse Conglomerates

Although it is generally agreed that some turbidity currents were capable of transporting pebbles and even small cobbles (Dott, 1963), geologists have been loathe to postulate that such currents could carry boulder-size clasts in suspension. Kuenen (1951, p.30) considered that boulders could only be rolled along the bottom by a turbidity current, and hence would lag behind and be by-passed by the flow: "to carry a large fragment right down to the end of a canyon, a considerable number of turbidity currents must occur one after the other."

In discussing the origin of coarse conglomerates in a flysch sequence, Natland and Kuenen (1951 p.104) concluded that turbidity currents were probably not responsible, "nor does it appear probable that the cobbles and boulders rolled down a submarine slope individually". They inferred an origin from landslides or sandy mudflows. Dott (1963, p.123) suggested that graded coarse conglomerates could result from down-slope rolling of the clasts (the large ones arrive first) while "any associated sand or mud could have moved in true suspension and have settled somewhat later onto the gravel".

In the present case, the complete mixing of boulders, pebbles and sand, and the presence of isolated boulders in the upper sandy parts of some beds (Plate 22B), strongly suggests that all the material was deposited more or less simultaneously. The boulders were not rolled into position and then covered by sand at some later stage. Transport of the boulders by some process which kept them above the bottom, at least for part of the time, seems inescapable.

The difficulty in imagining how boulders can be transported in sandy suspensions (not mudflows) arises because it is assumed that the normal fall velocity of the clasts must be overcome by turbulence. If, however, the fall velocity is significantly reduced because of some

other process which operates under conditions of high concentration and high velocity, then the problem may not be so difficult. Such a reduction in fall velocity is thought to occur in heavily-loaded antidune flows, and may be due to the dispersive pressure of grain collisions, as discussed below.

d. Transport by Antidune Flows

Although very little work has yet been done on sediment transport and deposition from flows in the upper part of the upper flow regime, some preliminary results are available. Nordin (1963) has shown that ephemeral streams in flood, which are characterized by standing waves and violently-breaking antidunes, may carry more than 60% by weight of sediment, three-quarters of which may be sand. This great capacity for sand transport is partly dependent on the presence of a significant amount of fine material, which increases the density of the fluid to such an extent that the fall velocity of the sand is considerably reduced. Nordin (*ibid*, p.19) states that "presumably, at sufficiently high concentrations of fine material, the fall velocities of the sand would be so reduced that the rate of sand transport would be similar to the transport of fine material; that is, the transport rate would be limited either by the availability of sand or, at the upper limit, by the concentration at which the sand particles are in actual contact."

The upper limit of such conditions would perhaps be a mud-flow, but the important point is that antidune flows can transport this amount of sediment in suspension and still retain considerable velocity (up to at least 6.48 feet per second according to Nordin). Transport similar to this (stream floods and sheet floods) are known to transport pebbles and boulders on alluvial fans (e.g. Blissenbach, 1954). Here, the driving force for the currents is largely provided by the slope, and the slope factor must also be critical for submarine flows.

At these high concentrations, Nordin is doubtful of the effectiveness of turbulence in supporting the sediment in suspension, and suggests that the dispersive pressure due to grain collisions, as demonstrated by Bagnold (1954), would be more important. This same

mechanism has already been applied to turbidity currents by Walker (1965) and Sanders (1965) in connection with the deposition of the basal structureless division of graded beds. Kuenen (1951), however, was of the opinion that turbulent flow would still be possible in flows of these concentrations, and stated (*ibid*, p.24) that "sand suspensions with densities of 1.8 can form turbidity currents, and with a suitable combination of clay, sand and gravel, the highest limit is about 2.0". This corresponds to a volume concentration of about 0.6 and a weight concentration of the order of 80%, but turbulent flow was not demonstrated at these densities.

The phenomenon of reduction of fall velocity in concentrations above about 1% is well known (e.g. Briggs and Middleton, 1965, p.11), although its effect on laboratory flume experiments can generally be ignored. In very dense flows, however, it must be critical (Middleton, 1966, suggests 78% reduction at 40% concentration), and possibly offers a mechanism by which pebbles and boulders can be transported in suspension. Bagnold (1962) and Middleton (1966) have emphasized that a considerable reduction in fall velocity is probably a necessary criterion for auto-suspension in turbidity currents.

Any hypothesis based on an analogy between subaerial stream flows and submarine density flows must necessarily be very tentative, however, because the surrounding media are so different. The overlying seawater would obviously exert a considerable damping influence on any submarine flow because of the friction and mixing. Some of the theoretical treatments of turbidity current flow have ignored this effect (e.g. Stoneley, 1957; Bagnold, 1962), but Middleton (1966) considers that the mixing and resistance at the upper interface at Froude numbers greater than one (i.e. upper flow regime) would be too great for steady flow (i.e. auto-suspension) to exist. He suggests, therefore, that coarse turbidites result from currents which did not achieve auto-suspension but which were depositing sediment at all stages. Evidence has already been presented, however, that gravel-bearing currents passed over parts of the seafloor without depositing, a fact which implies auto-suspension. The boulder-bearing flows, however, may not have achieved this condition.

G. RECONSTRUCTED NATURE OF THE DEPOSITING CURRENTS

The preceding analysis of the bedded rocks of the Singing Creek Formation strongly suggests that the conglomerates, sandstones and siltstones were all deposited by the one type of current, the lateral and longitudinal variations within the current causing different grain-sizes to be deposited in different places. The conglomerates (Group D) and thick-bedded sandstones (Group C) were apparently deposited mainly by the initial part of the current, in which there was little sorting of the sediment or production of lamination during deposition, while the thin-bedded sandstones (Group B) were deposited mainly by the lamination-forming phases which followed this initial phase. The siltstones (Group A) apparently represent the final stages of deposition from the tail-end and margins of the current.

The currents were: (i) spasmodic and waning, as indicated by the graded bedding and the sequence of internal structures; (ii) bottom-hugging, as indicated by the influence of bottom features on deposition, and by the presence of flute marks etc.; (iii) capable of carrying a mixture of silt, sand, pebbles, and in some cases even boulders, in suspension; (iv) probably in the high upper flow regime as deposition began, as indicated by the sequence of internal structures and by the occurrence in some beds of probable antidune lamination (dish structure). The presence of isolated coarse-grained depression-fillings at the base of many beds indicates that the initial gravel-bearing phase of the current was capable of passing over areas without depositing (except in the hollows), a condition approaching auto-suspension. Transport of boulders by some flows, and of pebbles by many flows, strongly suggests that some process other than turbulence was responsible for maintaining the coarse material in suspension, and Bagnold's dispersive pressure seems the only mechanism capable of this.

The following reconstruction combines the important features deduced from the present study. The currents were sediment-laden density underflows in which the bulk of the coarse material was carried in a dense frontal and basal part where flow conditions were in the high upper flow regime (antidunes and above). The dispersive pressure created by grain collisions in this fast-moving frontal part was

probably largely responsible for maintaining the mixture of silt, sand, pebbles and even boulders in suspension. Deposition from this part was strongly influenced by bottom topography, and tended to occur in depressions. The bulk of the material was deposited very rapidly, and generally over a fairly restricted area, producing thick beds (Groups C, D). Antidune lamination (dish structure) was formed in some cases when the sediment load consisted mainly of fine sand, but otherwise the deposit tended to be structureless and unsorted (division a). Isolated small remnants of this material were commonly left in scour marks or depressions in the areas up-current from the main deposit.

Deposition was slower as the current velocity declined to the plane bed phase of the upper flow regime, allowing the formation of a thin moving bed load wherein tractional and sorting processes could operate to produce parallel lamination. Some re-working and smoothing-off of the previously-deposited sediment occurred during this phase. Most of the gravel and coarse sand had already been dropped, and the combination of finer grain size, lamination, and better sorting (including removal of much of the fine material) often resulted in this division (b) having a sharp contact with the underlying division a. Because there is only a narrow range of flow power appropriate to dunes or mega-ripples in the fine sands which remained at this stage, the plane bed was usually replaced directly by ripples as the current declined rapidly through the lower flow regime. Deposition from these phases of the current tended to be more widespread than the early part and less influenced by bottom topography, producing thin laminated sandstone layers (many Group B beds) as well as laminated tops to the earlier deposits. In many cases this laminated sand was laid down directly on the flute-marked surface which had been scoured but left uncovered by the frontal part of the current.

The denser initial part of the current was flanked and followed by a more diffuse part in which mainly silt and fine sand were carried, probably in turbulent suspension. Deposition from this part tended to be blanket-like and widespread (Group A siltstones). Much of the silt deposited on the initial sand and gravel deposits was apparently eroded and incorporated into later flows following the same path, so

that the main areas of silt accumulation tended to be marginal to the coarse deposits. In some cases, silt from this phase was deposited directly on small remnants of gravel or even isolated pebbles left by the front of the current. Current velocities during deposition of the fine material were mostly in the lower part of the lower flow regime, involving formation of graded laminae possibly as large slow eddies impinged on the bottom. Fine-scale rippling of the fine sand also occurred. The last stages of deposition probably involved the settling out of the finest material from suspension.

Many of the flows occurred in rapid succession, probably before deposition from the preceding flow had ceased, and the silty tops of the beds were either not deposited or were eroded off so that composite sandstone sequences were formed. In other cases, a single current appears to have consisted of several surges of sub-equal velocity, each capable of eroding the deposits of the earlier one to produce scoured intra-bed surfaces. The high velocity which must have been attained by the flows, in order for them to reach the upper flow regime state and to overcome the friction and mixing effects with the surrounding seawater, was presumably induced by the steep slopes on which the currents were initiated and over which they travelled for most of their journey.

H. GROUP E - SLUMP SHEETS

1. General

Zones of deformed and mixed sediments are very common in the Singing Creek Formation, and constitute the bulk of the section in some areas (e.g. Fig. 11). The zones show abundant evidence of lateral movement, such as stretching, breaking, folding and rolling-up of beds, slurring, disaggregation and mixing, development of "flow cleavage" etc., and are confidently interpreted as being due to mass down-slope movement of unconsolidated sediments. They are termed slump sheets (after Kuenen, 1948) because the majority have sub-parallel upper and lower surfaces across the outcrop width (up to 200 feet or 60 m). The lack of correlation of slump sheets in outcrops more than a few hundred feet apart suggests, however, that many are lens-shaped rather than truly

sheet-like. The majority consist mainly of siltstone (Group A) with some intermixed sandstone and conglomerate, and all involve a number of beds. Practically every example is truncated at the top, usually by a sandstone bed, indicating that the slumps occurred at the seafloor prior to burial, i.e. they are open-cast.

The slump sheets range in thickness from a few inches up to about 20 feet (6 m), but in the measured section on the western flank of Flagstone Knoll (Fig. 11) the majority are 2-6 feet (0.6-2 m) thick. Zones up to 20 feet thick can be shown to consist of several superimposed slump sheets in some cases, but single units of this order of thickness also occur. The best exposures of slump sheets are on the upper western flank of Flagstone Knoll (Fig. 5), where a complete section through 320 feet (96 m) of intercalated slump sheets and sandstones has been measured (Fig. 11). The slump sheets within this section have been given reference numbers (1 to 40), some of which are referred to in the following discussion. The reference numbers are shown on Fig. 11. Although most of the following account is based on observations in this area, it is apparent that slump sheets of similar type and size are abundant within the formation in practically all areas.

2. Stages of Deformation

Examination of numerous slump sheets indicates that there is a complete spectrum of deformation beginning with stretching and simple folding and passing through complex folding, disruption and partial disaggregation to complete disaggregation and mixing, the end result being a completely reconstituted sediment. For the purposes of description it is possible to subdivide this spectrum into four overlapping stages, termed coherent, semi-coherent, incoherent and homogeneous. The major distinction lies between the semi-coherent and incoherent stages, marking the point at which viscous fluid flow became predominant over plastic flow (Dott, 1963). Some slump sheets show several different stages of deformation when they are traced laterally, but the majority are more or less uniform across the outcrop width. Whether the stage of deformation reflects the actual amount of lateral movement undergone by the slump is difficult to say in some cases, since much of

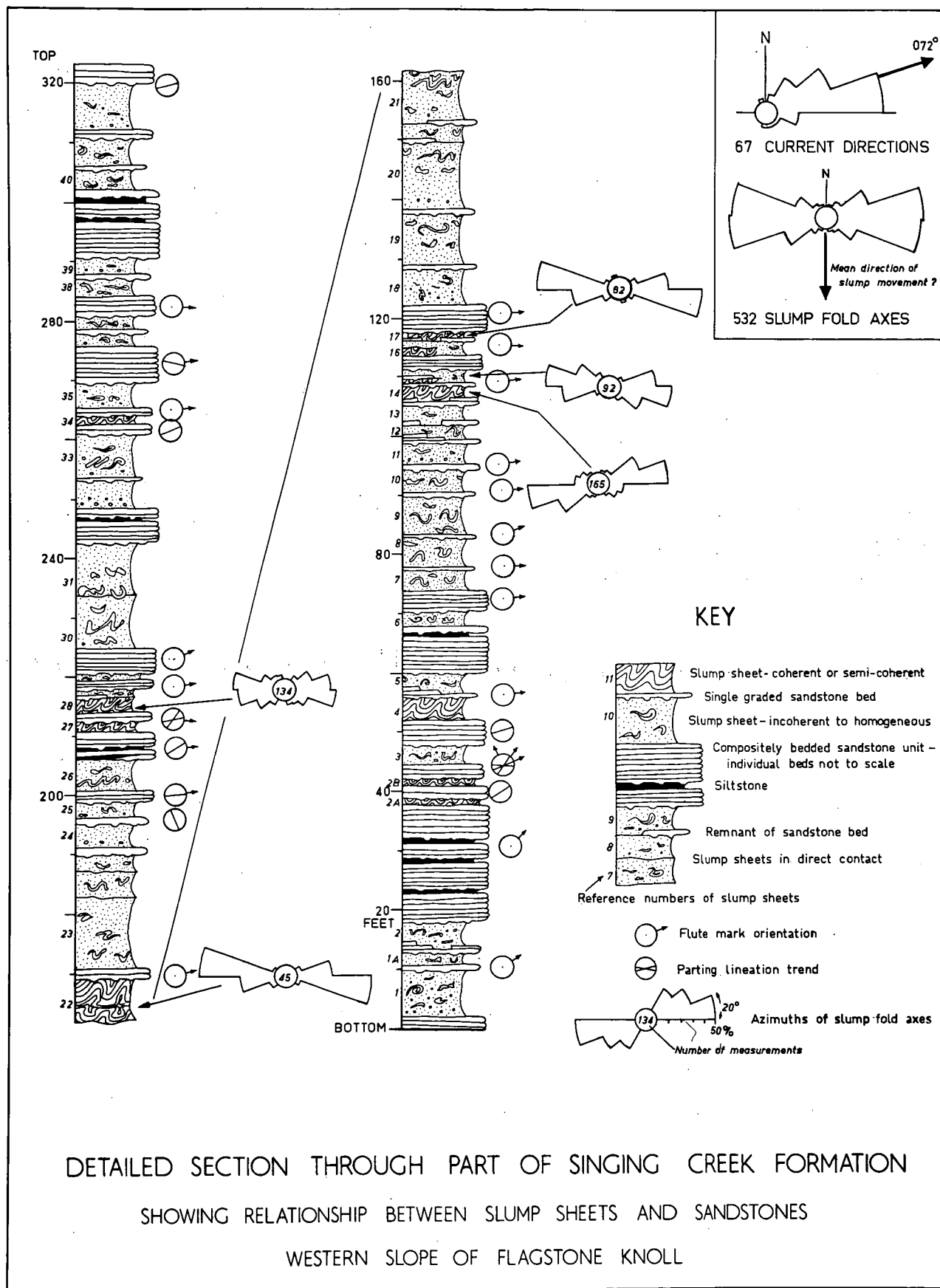


FIGURE 11

the movement may have been accomplished by sliding on a fluid decollement zone, with little deformation in the remainder of the slump. Parts of some slump sheets appear to have reached the incoherent stage without becoming folded in the process, i.e. the plastic flow stage consisted mainly of sub-horizontal stretching.

a. Coherent Stage

The term coherent is applied when the original bedding in the slump sheet is well preserved and is more or less continuous except where truncated at the top. The movement in such slumps has been almost wholly by plastic flow (Dott, 1963). Simple to complex folding (Plates 27, 28A), "necking" of sandstone layers, and pull-apart structures (Plate 28A) are characteristic of this type. Sub-horizontal bedding, usually showing "crinkling" of sandstone layers (e.g. Plate 10A), may also occur.

Examples of slump sheets showing coherent deformation are relatively rare in the Flagstone Knoll section (Fig. 11), where the majority are either incoherent or homogeneous. This is probably a reflection of the poorly consolidated and saturated nature of the beds at the time of slumping, and of the fact that deformation occurred within the topmost few feet of superficial sediments with no confining load pressure.

b. Semi-coherent Stage

At the semi-coherent stage much of the original bedding is preserved but there is also a significant proportion of slurried or disaggregated sediment and many of the beds are not continuous for more than a few feet because of disruption, disaggregation or attenuation. The folding is usually complex and irregular, and the folded layers may be contained in a "matrix" of disaggregated sediment (Plates 28B, 29A). The presence of reconstituted sediment in which the bedding has been destroyed indicates that the liquid limit was exceeded in parts of the slump, and that some of the movement was by viscous fluid flow.

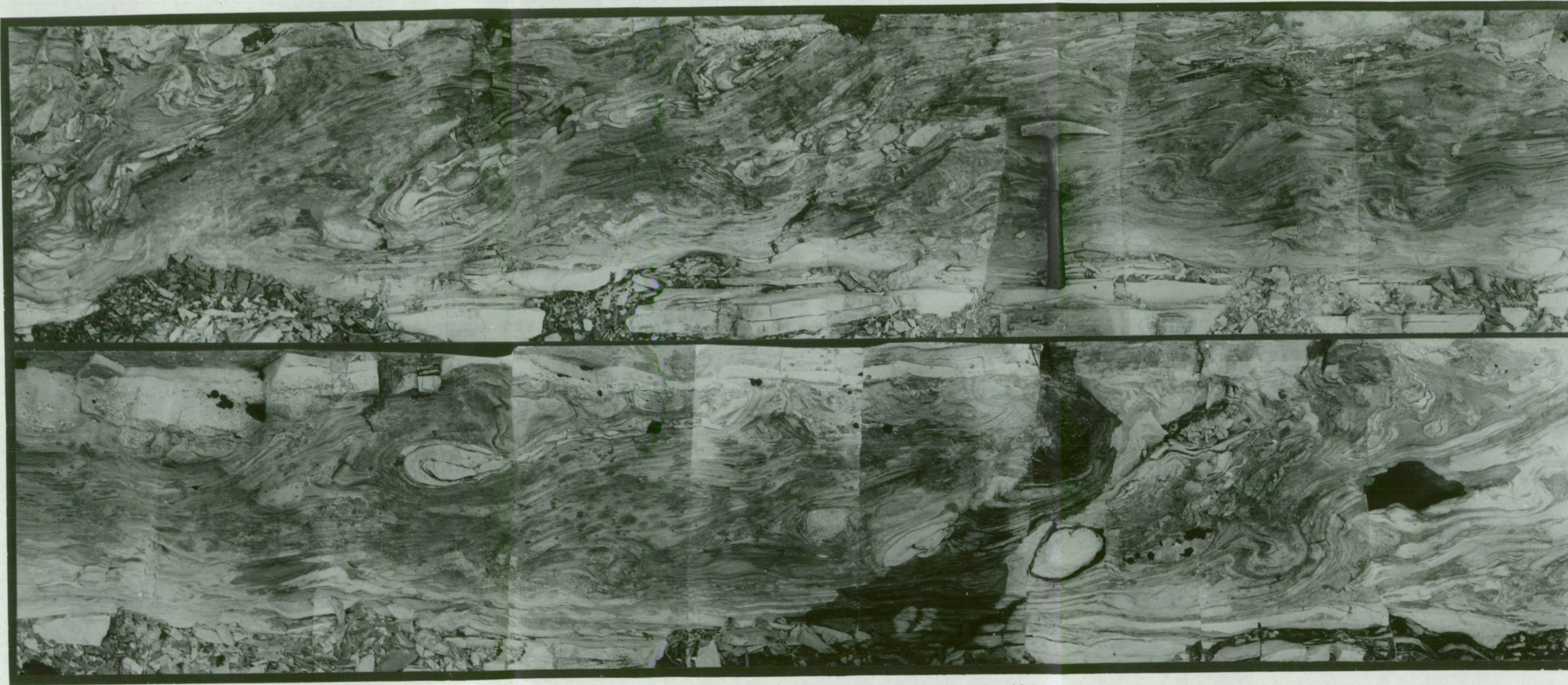


Photo series showing 20-foot section of Slump No. 14. Continuous from upper left (north) to bottom right (south). Fold pattern consists of sigmoidal nappe-like anticlines of silty sediment, showing incipient flow cleavage, separated by very compressed synclines in sandy layers. Some of the synclines consist of non-coaxial folds pressed tightly on top of one another, e.g. those to left of hammer. These synclines also show small parasitic folds. Folding dies out downwards by means of several decollement zones. Note truncation of slump sheet by graded conglomeratic bed which fills irregular depressions in top of slump. Only small remnants of this bed are preserved towards the northern end where the overlying slump sheet is in direct contact with Slump 14. Note very attenuated flap-like folds at southern end.

SOUTH

c. Incoherent Stage

In most of the slump sheets the original bedding is preserved only as remnants "floating" in a matrix of disaggregated sediment, and this stage is termed incoherent. The bedding remnants usually consist of sandstone in the form of irregular balled-up folds, lenses, wisps, or sub-angular blocks (Plates 29B, 30A). The reconstituted matrix consists in most cases of very poorly sorted sandy siltstone or silty sandstone. Scattered pebbles may also be present, and are abundant in some slumps (e.g. Plate 29B). Irregular patches of partly slurried sediment, in which faint bedding is still apparent, may also occur. Movement in the final stages of these slump sheets was apparently mainly by viscous fluid flow.

Flow cleavage

A feature which is best developed in some of the incoherent and homogeneous slumps, but which also occurs in parts of some semi-coherent slumps, is a weak secondary foliation which may be termed "flow cleavage". This structure occurs in disaggregated material in which the original bedding has been partly or completely destroyed by flow (e.g. Plate 30A, B; see also Plates 9A; 11A, B; 12A; 15A), and is a very prominent feature in some homogeneous slump sheets. The cleavage is seldom well developed or uniform, but is accentuated by weathering in many outcrops. Mesoscopic observation suggests the cleavage is due to orientation of minerals, particularly micas, but it has not been studied in thin section. Inter-reaction between cleavage and fold crests produces a prominent lineation in many places, as discussed under Folding.

The orientation of the cleavage in incoherent slumps is seldom constant for more than a few tens of feet, and in many cases it shows large complex folds presumably inherited from folds in the original bedding. Sharp truncation of the flow cleavage by the sandstone bed above the slump sheet gives rise to impressive local unconformities in many places (e.g. Plates 11B; 12A), and demonstrates that the cleavage is a product of the slump movement and is not a later "tectonic" feature. Observations on coherent and semi-coherent slump



A. Complex coherent deformation in part of Slump No. 14. Note truncation by overlying sandstone; complexly refolded folds; downward decrease in deformation; pull-apart structure in sandstone layer at base.



B. Semi-coherent deformation in Slump No. 17. Folded layers in matrix of disaggregated sediment. Slump lineation is well developed. Note alignment of folds perpendicular to outcrop surface and parallel to current marks on underlying sandstone bed.

sheets suggest that the cleavage develops in areas of maximum movement or flow, in which the bedding is gradually obliterated by slurring to be replaced by the sub-parallel flow lines. Slurring and incipient cleavage development are apparent in some of the lobe-like anticlines in Slump 14 (Plate 27).

d. Homogeneous Stage

Some slump sheets consist almost entirely of reconstituted sediment, with only scattered isolated remnants of original beds (usually sandstone "balls" or blocks), and this stage is termed homogeneous. The sediment is usually a very poorly sorted gritty micaceous sandstone with scattered pebbles, but may be very pebbly or predominantly fine-grained. In some cases there are considerable variations in sediment type (pebble or sand content etc.) both laterally and vertically, and the term homogeneous is meant to indicate the lack of original bedding rather than true homogeneity of composition.

Two features are more or less characteristic of this stage of deformation. The first is the fact that the flow cleavage when present, tends to be sub-horizontal throughout the slump, with only gentle undulations (Plate 30B). This probably reflects the more fluid nature of the slumping material, with the flow tending to become laminar as the internal inhomogeneities are destroyed.

The second feature is a tendency towards a vertical size grading within the slump. The lower part of many slumps of this type tends to be predominantly sandy or pebbly whereas the upper part is more shaly and finer-grained, with few pebbles. While the prevalence of this feature (17 of the slump sheets in the measured section of Fig. 11 show crude grading) suggests it may have resulted from the settling of coarser particles during flow, another explanation is possible in some cases. This is that the size grading is inherited from the original pre-slump sequence, which may have been dominated by sandstones in the lower part and by siltstones in the upper part, only limited mixing having taken place during the flow. Sequences of this type are preserved in a few semi-coherent slumps. As well as this, the pebbly layer which occurs at the base of some homogeneous slumps



A. Semi-coherent deformation in Slump No.28, with large synclines and complex anticlines. Contrast in deformation between upper and lower parts suggests two interfering slump sheets involved. Note slurried contact on underlying sandstone.



B. Typical incoherent slump sheet with irregular sandstone remnants and "floating" pebbles.

(e.g. Plate 8A, B) can be shown with reasonable certainty to have been derived from a pebbly sandstone bed which is preserved at this level elsewhere. In most cases, however, the reason for the grading cannot be demonstrated.

3. Folding

a. General

The style of folding and the kinds of plastic deformation shown by the slump sheets are of some interest because of possible correlations with "tectonic" structures and because they give some indication of the nature of the sediments at the time of deformation, the reason for the slumping, and possibly the direction of movement. Only those slump sheets in which the deformation is coherent or semi-coherent are of much value in this respect. An analysis of the slump folding using the techniques of structural geology is beyond the scope of this thesis, and the following observations are intended only to give a general impression of the kinds of deformation present.

A study of the literature reveals some lack of agreement concerning which rock types have behaved competently and which incompetently during slump folding. In the present case the rock types involved are mainly thinly interbedded quartzose sandstones (fine- to medium-grained) and micaceous sandy siltstone, with some thicker sandstones and conglomeratic sandstones in some places. It is apparent that none of these rock types has been particularly competent, since all show marked thickness changes around folds and disaggregation in mobile zones. It is also apparent, however, that at least some of the thicker sandstone layers have been relatively more competent than the sandy siltstones, since it is the sandy beds which have best retained their identity during all deformation stages from coherent to incoherent and which are preserved as remnants in most of the slump sheets.

The folds are mostly very irregular and complex, and the pattern of folding varies between slumps and within slumps, such that generalizations are difficult. The folding has been predominantly by plastic flow, however, and the term "flow folding" is generally



A. Incoherent slump sheet (No.20) with flow cleavage, overlain by sandstone raft showing deformed flute mark. Note unusual pull-apart structure at left-hand end of sandstone layer; also lineation on fold within slump sheet.



B. Homogeneous slump sheet (No.2) with horizontal flow cleavage. Note disrupted sandstone beds at base.

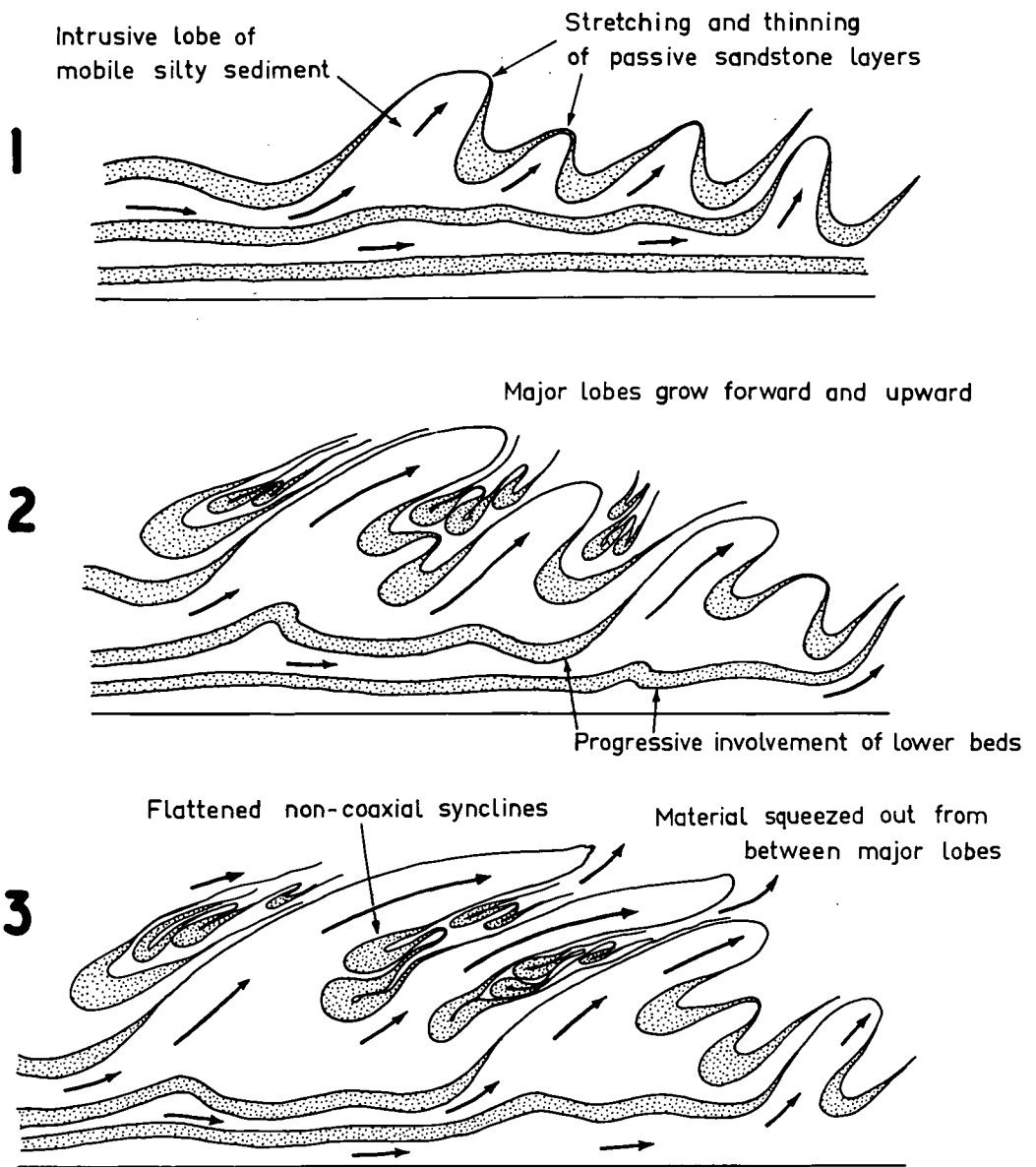
applicable. Open, concentric-type folds are exceptional. Clear-cut faults or other indications of brittle failure associated with the folding have not been observed, although incipient smeared-out glide planes occur in a few places. A number of types of folds are recognizable, some of which appear to correspond to types which are well known in structural geology. Some of the more common ones are: isoclinal folds, refolded or irregular folds, box-type folds, broad synclines associated with complex anticlines, and parasitic folds.

b. Isoclinal folds

Those folds which are not irregular or rolled are nearly all isoclinal, with elongate flattened limbs and tight hinges (e.g. Plates-27; 30A). These folds seldom form a regular pattern, most of them being sandwiched within more complex structures or more or less isolated in the slurried matrix of incoherent slumps. Refolding of isoclinal folds to form secondary folds which are themselves isoclinal is a common feature.

An example of a fairly regular pattern of isoclinal folds, in which the mode of development of the folds can be postulated with some confidence, occurs in Slump No. 14 of Fig. 11 and is shown in Plate 27. This pattern persists over a strike interval of about 25 feet (8 m), passing into less regular deformation on either side (e.g. Plate 28A). The pattern is characterized by very attenuated folds in which the axial surfaces dip north and tend to be sigmoidal, with sub-horizontal upper and lower portions. The anticlines are developed mainly in siltstones and fine sandstones, and tend to over-ride one another like small-scale nappes. The synclinal areas are dominated by thicker sandstone layers forming very tight flap-like folds which are markedly attenuated on the limbs. In some cases the synclines are represented only by teardrop-shaped remnants of sandstone.

Several features of this fold pattern are significant. When the lower, relatively undisturbed beds are traced to the right (south) it is found that they become progressively involved in the deformation, i.e. the slump sheet is incorporating successively deeper horizons to the south. The thick sandstone which forms the base of the



SUGGESTED DEVELOPMENT OF ISOCLINAL FOLD
PATTERN IN SLUMP NO. 14

FIGURE 12

slump throughout the illustrated section becomes involved in the folding just beyond the photographed area. These progressively-involved horizons must have been separated from the upper folded zone by decollement zones or surfaces, as shown diagrammatically in Fig. 12.

The anticlinal folds are characterized by poorly defined bedding, slurring, and incipient flow cleavage development, and appear to have been zones of considerable sediment movement accompanied by stretching and disaggregation of beds. They resemble diapiric lobes which have grown upwards and forwards through the overlying sediment. The synclines, on the other hand, appear to have been relatively passive, and it is only in the noses of these folds that the beds do not show marked thinning or disaggregation. It is notable that the very tight folds which make up the major synclinal zones are not all co-axial, i.e. the synclines consist in part of a series of flattened folds, several of which may belong to the same bed, pressed tightly on top of one another. The sediment between the sandstone layers forming these flattened synclines must have been squeezed out when the fold limbs were compressed together (Fig. 12).

The fold pattern appears to be the result of the upward and forward (down-slope) growth of lobe-like anticlines, composed of mobilized silty sediment derived from the lower parts of the slumping sequence (Fig. 12). This material has forced its way through the overlying "passive" sandstone layers, causing stretching and attenuation of these layers except in the synclinal cores. Collapse of the major anticlinal lobes onto one another, and consequent squeezing-out of the silt from the zones in between, has resulted in the cluster of tightly-pressed folds which make up the synclinal areas. The complex, partly-slurried zone at the top of the slump is probably a mixture of superficial sediment and material extruded from below. The reason for the relative mobility of the silty sediment forming the anticlines is not so apparent, but could be due to a combination of liquefaction and gravity loading by the overlying material.

c. Irregular Folds

The majority of folds are too irregular to be defined according to their geometry. Many appear to be isoclinal folds which have been refolded several times (e.g. Plate 28A), or rolled up to form complex balled structures. The shape of these rolled structures in three dimensions is difficult to gather from the exposures, but at least some appear to be almost spherical.

d. Box-Type Folds

Folds which approach the form of box folds or kink folds are common in Slump No. 22 (Plate 31A), and there are isolated examples in other slumps. These folds range in size from an inch or so high to several feet, and may be simple or conjugate. They seem to occur only in thinly bedded sandstones and siltstones, and thick sandstone layers have not been observed to form folds of this type. The double hinges of the folds are commonly somewhat rounded, such that some of the folds approach a semi-circular form.

e. Broad Synclines and Complex Anticlines

Some slumps are characterized by a pattern of alternating broad synclines and complex anticlines (Plate 29A). The synclines usually consist of relatively thick-bedded sandstone and are only weakly deformed, while the anticlines may show complex deformation and disaggregation, and usually consist of relatively thin-bedded material originally underlying the sandstone. The synclines may contain a number of beds, in which case sedimentary structures such as graded bedding, sole marks etc. may be preserved (e.g. Plate 29A), or they may consist of a single bed, in which case they usually appear as massive lenses or balls of sandstone (Plate 7B). The contrast in deformation strongly suggests that the anticlines have been mobile and have intruded through the relatively passive overlying beds in much the same manner as that just described for the isoclinal folds. In some cases, a similar pattern has resulted from the deposition of a sandstone bed on top of an already-formed slump sheet, and is due largely to loading rather than to lateral movement.



A. Large conjugate box-type fold in semi-coherent slump sheet (No. 22). Note small parasitic folds on sandstone layers within major fold.



B. Sandstone rafts folded into top of slump sheet (No. 16) and truncated by overlying sandstone bed. Overlain by semi-coherent slump sheet.

f. Parasitic Folds and Slump Lineation

Many of the large folds have associated with them a characteristic small-scale crinkling or crenulation of thin sandstone layers (Plates 27; 28B; 29A; 30A; 31A). The crenulations may affect the whole bed, in which case they form a series of small tight folds (Plates 27; 31A), or they may occur on one side only (Plate 28B). In the latter case they appear as a series of linear, parallel steps on the bed surfaces, forming a prominent lineation. The crenulations are always parallel to the major fold axes when they can be seen in three dimensions (e.g. Plates 28B; 30A). They appear to be similar to the parasitic folds or "drag folds" well known in structural geology (e.g. Whitten, 1966) and, in some cases at least, their relationship to the major fold is the same as that of drag folds. The lineation is apparently due to the intersection of flow cleavage with the bedding, causing slight displacement of the bed surfaces.

4. Basal Contacts

The basal contacts of the slump sheets are of two main types, viz. (i) sharp contacts along more or less flat surfaces parallel to bedding (i.e. décollement surfaces), and (ii) gradational contacts wherein the deformation decreases downwards. The latter type, however, generally involves one or several décollement zones. Sharp contacts are characteristic of incoherent and homogeneous slump sheets, while coherent and semi-coherent types generally have gradational contacts, as might be expected.

Most contacts are of the single sharp décollement type, with the reconstituted slump material resting abruptly on an undisturbed sandstone layer. The upper fine-grained part of this sandstone bed, if it existed, has been incorporated into the slump sheet in virtually every case. Very commonly, this sandstone bed rests directly on an underlying slump sheet, and is the only surviving remnant of the original bedded sequence (see Fig. 11). The reason for its survival is not always apparent, but in many cases it is probably due to the bed being "anchored" to the top of the underlying slump sheet by means of the sole marks. Most beds in this situation have sole marks which penetrate

deeply into the underlying slump material (e.g. Plates 9A; 11A; 13A, B; 28B; 30A), and if this material had become at all cohesive by the time the overlying later slump occurred, then the sole marks would tend to hold the lower part of the sandstone bed in place and prevent it being incorporated in the movement. The sole marks are all that remain of these beds in some places (e.g. bed above Slump No. 14 in Plate 27).

The décollement surface commonly shows minor irregularities due to slurring or to local removal of parts of the sandstone bed (e.g. Plates 29A; 8B). In rare cases the middle levels of the bed have been slurried while the upper part remained more or less intact (e.g. Plate 10A). Larger-scale irregularities, particularly those due to progressive removal of deeper beds, are also common (e.g. Plate 8A, B). In a number of places the complete bedded sequence has been incorporated into the overlying slump sheet, such that the two slump sheets are in direct contact (e.g. Plate 27; Fig. 11). Such contacts, when recognizable, are generally sub-horizontal, but large sandstone folds or "balls" in the upper slump may penetrate deeply into the lower one. The example shown in Plate 29A may be of this type. These irregular contacts probably grade into indistinguishable contacts with further mixing, and several examples of complete amalgamation of two slump sheets to form a single thicker (and less coherent) unit have been observed.

Gradational basal contacts are less common, and usually involve the partial infolding of sandstone layers near the base of the slump, with associated development of minor décollement zones in the less competent siltstones. Pulling-apart or "necking" of sandstone layers is another feature of such contacts (e.g. Plate 28A), and has resulted in the formation of sandstone "rafts" in some places. The gaps between the rafts are filled with siltstone which has welled-up through the weakened part from below (Plate 28A). Such contacts are characteristic of coherent and semi-coherent slump sheets, but there are also a few examples of homogeneous slumps which have disrupted and partly incorporated some of the underlying beds to form a gradational contact (e.g. Plate 30B).

Many of the single sandstone beds which survive between two slump sheets show evidence of lateral movement at some stage after deposition. Such movement is indicated by deformed sole marks (e.g. Plates 13A; 30A), by slight folding of the bed, or by "rafting" (Plate 30A). The sole marks (particularly flute marks) appear to have been dragged sideways after their formation, resulting in very asymmetrical shapes and, in extreme cases, in the complete detachment of the structure from the bed. The example in Plate 30A shows a deformed flute mark with a series of small steps on the oversteepened left-hand side, apparently developed by progressive internal shearing as the bed moved from left to right. All the known examples of this kind of deformation occur on beds which are directly overlain by a slump sheet, strongly suggesting that the deformation was caused by the drag of this over-riding slump.

The sandstone layer shown in Plate 30A is actually a raft about 10 feet long, and shows an unusual pull-apart structure at its left-hand end. Instead of the usual necking or diapiric structure, the break has apparently developed along several sub-planar surfaces which may have been compactional joints. Slight tensional opening is apparent on one of these joints adjacent to the main break.

5. Upper Contacts

Most of the upper contacts of the slump sheets are abrupt and erosional, with a horizontal sandstone bed truncating the deformed bedding or flow cleavage of the slump. Thus most of the slumps must have occurred at the seafloor prior to deposition of the overlying bed. Some contacts, however, suggest that the slump was rejuvenated after deposition of some overlying sediment. Thus a bed or sequence of beds which truncates the slump may itself be deformed and then truncated by an overlying bed (e.g. Plate 31B). Usually, such beds or sequences appear only as isolated rafts along the top of the slump sheet, since only the more deeply infolded parts have been preserved. The more extreme examples of this type could be regarded as representing two separate slump sheets (e.g. Plate 29A).

6. Directional Properties

Consideration of the directional properties of slump sheets involves two closely related factors, viz. the azimuth of the movement, and the sense of the movement, and while both of these can be determined from the same criteria in some slumps, different criteria are involved in many cases.

a. Azimuth of Movement - Fold Axis Orientation

Two kinds of criteria have been used by previous authors to determine the azimuth of slump movement, viz. (i) statistically oriented linear or planar elements within the slump, and (ii) linear features on the base of the slump. The internal elements used include fold axes, glide planes, slabs of competent rock, and pebbles or larger clasts. The sole marks used include slide casts, channels and striations. In the present case the only available criteria are the slump fold axes, since these are the only internal features present in sufficient quantity. The soles of the slump sheets are exposed only in section.

Some difference of opinion exists as to whether the orientation of slump fold axes reflects the real direction of movement and the palaeoslope. Dzulynski and Walton (1965, p.193) and Waterhouse and Bradley (1957, p.530) have argued that the local variations in direction within the main mass of the slump may be so pronounced as to obscure the main movement, but most authors have followed the original proposal of Jones (1939) that the fold axes will tend to be parallel to one another and to the strike of the palaeoslope down which the slump moved.

Many authors have found that slump fold axes are aligned, e.g. Jones, 1939 (over 100 strikes); Chenoweth, 1952 (18 axes); Newell et al., 1953 (30 axes); Rigby, 1958; Murphy and Schlanger, 1962 (92 axes); Marschalko, 1963; Scott, 1966 (202 axes); Rattigan, 1967 (55 axes), and Powell, 1967 (231 axial planes), although the statistical significance of the measurements is doubtful in some cases. Some of these authors found, however, that the direction of movement indicated by the slump folds was in conflict with the regional palaeoslope orientation deduced from other evidence such as facies changes and provenance. As well as this, there are many recorded examples of discordance between the

indicated slumping direction and the associated palaeocurrent direction, and this has led to arguments regarding the validity of slump fold measurements and also regarding the nature of the currents (e.g. Kuenen, 1967; Marschalko, 1963; Murphy and Schlanger, 1962). Some of this controversy seems resolvable if it is considered that the slope on which the slumps formed was a local feature not necessarily related to the regional slope down which the currents were travelling.

Fold axis orientation has been measured in five of the more coherent slump sheets of the Flagstone Knoll section, and the results are shown in Fig. 11 (p. 134). Initially, both the azimuth and dip of each fold axis were measured in two of the slump sheets, and were plotted on equal-area diagrams after rotation of the bedding (which dips 55° east) to horizontal. The axes were found to cluster about the horizontal plane, as might be expected, and it was considered that measurement of the dip of the axes was not necessary and did not warrant the extra time involved. Consequently, only the azimuths were recorded and, after unrolling of the bedding, these could be grouped mathematically and plotted as rose diagrams as shown. The five slump sheets treated in this way were selected for the abundance of fold axes preserved, since the majority of the slumps contain too few linear elements to give statistically significant information. The measurements were taken across the exposed widths of the slump sheets, involving a strike distance of 200-300 feet (60-90 m) in each case. The 45 measured folds in Slump No. 22 (Fig. 11) include a number of box-type folds.

Results

The fold axis plots indicate a remarkable degree of alignment in an east-west direction in each slump sheet, and a remarkable correspondence between the slump sheets. This strong alignment is perpendicular to the outcrop surfaces (which trend roughly north-south), and is apparent from even a casual inspection of the outcrops (e.g. Plates 27; 28B; 29A; 31A). That this alignment reflects the direction of movement, and is perpendicular to it, is strongly suggested by the illustrated part of Slump No. 14 (Plate 27), where the direction of movement has fairly obviously been from north to south (left to right) and the fold axes are strongly aligned east-west.

The direction of alignment is unusual because it indicates that the palaeoslope had an east-west strike, whereas the associated palaeocurrents indicate a palaeoslope with a roughly north-south trend parallel to the regional strike. The significance of this disparity is discussed later.

b. Sense of Movement

Criteria which have been used by previous authors to demonstrate the sense of movement of slump sheets include the following: dip of fold axes or of glide planes; direction of overturning of simple folds or of displacement on glides; imbrication of clasts, sediment blocks or "sandstone whirlballs"; direction of closure of isolated tear-drop-shaped folds ("hook folds"); direction of overturning of prised-up basement beds ("slump overfolds"). All of these are considered valid except the direction of closure of "hook folds", since there are several examples in the present slump sheets of adjacent folds which close in opposite directions.

Evidence for the sense of movement is lacking in most of the slump sheets examined, since the majority tend to be internally homogeneous and to have flat featureless soles. Unequivocal evidence is available in only two cases, viz. Slump No. 14, in which the overturned folds indicate movement to the south (Plate 27), and Slump No. 2A, in which a prised-up basement bed also indicates movement to the south. Strong evidence for southerly movement is available for three slumps (Nos. 11, 17, 21 of Fig. 11) which rest on sandstone beds showing dragged sole marks. Weak evidence is available in two other cases, viz. in Slump No. 33 several large sandstone slabs appear to be imbricated and suggest southerly movement, and in Slump No. 26 an isolated series of overturned folds again suggest southerly movement. Another feature which suggests movement towards the south is the progressive incorporation of underlying beds shown by some slump sheets as they are traced south, and the amalgamation of some adjacent slumps in this direction. Evidence for northerly movement has not been observed in any of the slumps in the Flagstone Knoll section, and it is concluded that the majority were formed on a south-dipping palaeoslope.

7. Genesis of the Slump Sheets

Crowell (1957) showed that "pebbly mudstone" slumps in the Ventura Basin had been initiated by the sudden deposition of gravel and sand from a turbidity current onto unconsolidated and possibly metastable muds. The gravel "sank as great lobes into the substratum", resulting in mixing of the two types of sediment accompanied by down-slope sliding, the end product being a pebbly mudstone. The trigger mechanism for the slumping in this case was the gravel-bearing turbidity current.

Such a mechanism does not seem to be generally applicable to the slump sheets at Flagstone Knoll, however. Most of the slumps, including the homogeneous pebbly types, show no evidence of having been initiated by deposition of a particular bed, and many do not contain any markedly coarse beds. Some, in fact, consist wholly of thinly-bedded siltstones and fine sandstones. Others contain conglomeratic beds underlain and overlain by fine-grained beds, all of which have slumped together. Several examples of deformation induced by the sudden dumping of sediment have been seen (e.g. Plate 25), but these are exceptional. The marked discordance between palaeocurrent direction and the direction of slumping also militates against current-induced slumping.

The slumps must have been initiated by some mechanism which rendered the upper few feet of sediments unstable over an area of at least several hundred square feet. In some cases at least this probably involved an increase in slope of the depositional surface, since the deformation dies out downwards, i.e. sudden liquefaction of a buried layer was not responsible. The frequency of the slumps, and the apparent constancy of movement direction, indicate that the underlying mechanism was probably a recurrent one with constant orientation. This suggests faulting, as discussed below.

8. Palaeoslope-Palaeocurrent Relationship

Current directions have been measured on many of the undisturbed sandstones which occur between the slump sheets in the measured section, including many of those which truncate the tops of slump sheets. Flute marks, which give the sense and direction of the

currents, and parting lineation, which gives the azimuth only, have both been measured, and the results are plotted on Fig. 11. In most cases it can be assumed that the parting lineation trend has the same sense as the adjacent flute marks, since the two are always sub-parallel when they occur on the same bed. Current directions from 67 beds on Flagstone Knoll are compared with the orientation of all measured slump fold axes in the inset of Fig. 11 (the 532 fold axes include 14 measured from Slump No. 2B).

The current directions lie predominantly within a 40 degree arc, with a vector mean of 072°T , i.e. the currents were mostly travelling east or north-east. This is almost parallel to the mean slump fold orientation (088°T), which means that the currents moved at right angles to the indicated direction of slump movement. This parallelism of slump folds and current structures is a striking feature of many of the outcrops (e.g. Plate 28B).

Several arguments suggest that the easterly direction of the palaeocurrents reflects the regional palaeoslope, viz. (i) this palaeocurrent direction persists through the overlying formations (see Fig. 17 of Chapter 4); (ii) the source rocks for the sediments lie to the west (Tyennan Geanticline); and (iii) the regional strike is north-south. There is no palaeogeographic evidence to suggest a regional palaeoslope oriented east-west and dipping south, as is indicated by the slump sheets. It is concluded, therefore, that the slope on which the slump sheets formed was probably a local feature which, although it persisted for some time, was not a regional slope controlling the course of the depositing currents. The most likely reason for such a local slope is faulting, and cross-faults which could have been responsible are present in the area.

The importance of contemporaneous seafloor faulting as a cause for the slumps which occur in many flysch sequences is becoming increasingly apparent (e.g. Gregory, 1969; Leitch and Mayer, 1969; Kuenen, 1967; McBride and Kimberly, 1963; Waterhouse and Bradley, 1957). The orientation of the faults may or may not be parallel with the regional trend, and hence the direction of travel of the slumps will not necessarily reflect the first-order palaeoslope. The slumps may

result either from tilting of the seafloor, causing over-steepening of the beds, or from the formation of contemporaneous fault scarps from which the sediments slide into the adjacent lower area, continually smoothing-off the topography. Earthquakes accompanying the faulting probably act as a trigger mechanism to initiate movement in metastable sediments.

The right-angle discordance between current direction and apparent slump-movement at Flagstone Knoll indicates that any faulting responsible must have been oriented at a high angle to the regional north-south trend. Preliminary analysis of the regional fracture pattern (see Chapter 1) suggests that the major north-south faults are connected by a series of northwest and east-west trending cross-faults. The Cambrian deposition seems to have occurred largely within tensional basins formed by these cross-faults. Thus the slump sheets could well be the result of seafloor movements on the cross-faults during sedimentation, and there is some local evidence for this.

At least two east-west trending minor faults affect the sequence at Flagstone Knoll (Fig. 5, p.45). One of these cuts through the crest of the knoll, and appears to have a south-side-down displacement. The fault is steeply inclined, and traverses the west flank of the knoll about 500 feet north of the measured section. Unfortunately there is little outcrop in this area, and the relationship between the slump sheets and the fault cannot be seen. The fault appears to die out upwards through the section, and cannot be shown to affect the overlying Reeds Conglomerate or Florentine Group. Thus it is probably an Upper Cambrian structure which was buried by later sedimentation. The south-side-down displacement would have created a south-facing scarp, and hence any fault-scarp slumping would have been directed towards the south. Those slumps for which evidence is available appear to have moved in this direction, as noted previously.

A second small fault is poorly exposed in Singing Creek on the south flank of the knoll. This fault is sub-vertical and appears to have a north-side-down displacement of about 100 feet. Again, it cannot be traced into the overlying Ordovician rocks. The faults are difficult to recognize because of the patchy outcrops and complex stratigraphy, and there are probably many others in the area.

I. DEPOSITIONAL ENVIRONMENT OF THE SINGING CREEK FORMATION

1. Major Features of the Deposit

The Singing Creek Formation is marine, as indicated by the fossils. The apparent absence of sedimentary features formed by waves or tidal currents, and the predominance of structures suggesting transport by density underflows (turbidity currents), strongly suggests that the formation was deposited below wave base and probably not on a continental shelf area. The conglomeratic nature of many of the beds indicates high-velocity currents, and implies proximity to the source and to a steep slope. The faunal assemblage includes a surprisingly high proportion of inarticulate brachiopods for a deposit of this kind. While many of these occur in the sandstone beds and have obviously been transported, there are also many in the siltstones, where they occur with trilobites and numerous trace fossils. It is just possible that these inarticulates have also been transported, since the bulk of the siltstone also appears to be of density current origin, but their abundance implies that they are partly autochthonous. The significance of the abundant inarticulates is not known, but it possibly suggests that the water depth was not great.

Evidence has already been presented that nearly all the sediments were deposited by the same general mechanism, viz. sediment-laden density underflows. Many of these flows were sufficiently powerful to carry pebbles, and some carried boulders. The anatomy of the deposit is dominated by the segregation of the sandstones and conglomerates into large lens-like bodies separated by sequences consisting mainly of siltstone with some interbedded thin sandstones. The three-dimensional form of the coarse-grained zones is unfortunately not known in detail, and remains a problem for future investigation. Consideration of similar deposits from various parts of the world strongly suggests, however, that the coarse-grained bodies represent channels or depression-fillings on a submarine fan complex, as discussed below.

2. Ancient and Modern Analogues

Although many flysch-like sequences are characterized by uniform, laterally-persistent graded beds and by apparent uniformity of

rock types over wide areas, it is becoming increasingly apparent that there is another type of flysch (usually containing conglomerates) in which marked lateral variations are the rule rather than the exception. Complementary to this, the study of modern ocean basins and deep-sea topography has shown that much of the deposition from turbidity currents takes place on submarine fans or aprons along the base of continental slope areas, and that within these fans there may be an irregular distribution of sediment types reflecting the irregular topography of the fan surface. Characteristically, this takes the form of a lateral differentiation into coarse-grained channel deposits and fine-grained inter-channel deposits. The realization that coarse-grained laterally-variable flysch sequences probably represent ancient submarine fan deposits constitutes an important advance in our understanding of flysch deposition.

One of the earliest interpretations of this kind was made by Passega (1953), who described an oil-bearing Tertiary turbidite sequence in which a lenticular arrangement of sandstones within a "matrix" of shale, siltstone and fine sandstone ("blanket beds") had been delineated by drilling. The sand lenses ("Colorado sand") have an average thickness of about 25 feet and a maximum of about 100 feet, while their lateral dimensions range from a few hundred feet to several miles. Each lens contains a number of graded sandstone beds, but correlation within lenses is difficult because of thickness changes and horizontal grading. The lenses can be correlated with sandy layers in the adjacent blanket beds, and seem to be "a local anomalous form of these beds". Electric log data indicate that the lenses are channels which have been eroded in the underlying blanket beds. Channels of the same age may occur several miles apart, and some areas (structurally-controlled ?) seem to be favoured by channels while adjacent areas consist almost wholly of blanket beds.

Passega postulated that both the channel sands and the blanket beds were deposited by turbidity currents on a fan-like area at the base of a submarine slope. The strongest flows eroded channels which were then filled with the coarsest material from later flows, while the finer sediment was spread outwards beyond the channel. This interpretation, made before any of the present-day fans had been investigated in

detail, agrees remarkably well with the results of the modern oceanographic studies.

The best known examples of modern submarine fans are those forming in a series of small structural basins off the coast of California. These fans have been described by a number of workers, including Dill et al., 1954; Menard, 1955, 1960; Gorsline and Emery, 1959; Emery, 1960a, b; Shepard and Einsele, 1962; Bouma and Shepard, 1964; Hand and Emery, 1964. The fans are delta-like bodies of sediment spreading outwards from the mouths of submarine canyons and in some cases coalescing to form compound aprons at the base of the slope. They show some remarkable analogies with subaerial alluvial fans, although they lie some 2000-6000 feet below sealevel.

Most of the fans are crossed by channels or fan-valleys issuing from the mouths of the canyons and sometimes anastomosing as they traverse the fans. Many of the channels have levee banks on either side and some show meanders and appear to have migrated laterally in the manner of rivers (Shepard and Einsele, 1962). The channels range from 100 metres or so wide and 10 metres deep on the upper parts of the fans to several kilometres wide on the lower parts. In some cases the present channels traverse the complete length of the fans and continue onto the basin floors, indicating that sediment from the canyons is now largely bypassing the fans and being deposited beyond them. Deep dissection of some fans by the present channels, leaving terrace remnants on either side, is apparent in some areas (Hand and Emery, 1964). Thickening of the fans on the downthrown side of contemporaneous faults, and non-depositional "shadows" behind obstacles (basement ridges etc.) are also reported, and demonstrate the gravity-controlled nature of the depositing currents.

The coarsest sediments (sand and gravel) are found in the channels, and include graded layers as well as massive and multiple beds. Ungraded sand and gravel layers are reported as being more common than graded layers in the channels by Bouma and Shepard (1964). The levees and inter-channel areas are characterized by "relatively low sand content, thin bedding, and large amounts of turbidite silt" (Hand and Emery, 1964, p.535). Moore (in Buffington et al., 1967) has

emphasized that nearly all the fine material which forms the bulk of many submarine fans is probably also of turbidity current origin. Several authors report difficulty in correlating layers between cores taken on the fans (e.g. Shepard and Einsele, 1962; Gorsline and Emery, 1959), and have suggested that many of the layers are laterally variable or discontinuous.

The mechanism proposed for these features is as follows (from Menard and Ludwick, 1951; Menard, 1955; Gorsline and Emery, 1959; Hand and Emery, 1964). Turbidity currents generated in the submarine canyons (by mass movement of sediment from the head of the canyon) are slowed down by the decrease in slope at the mouth of the canyon and begin to deposit sediment soon afterwards if the velocity-slope relationship is suitable. The existence of channels indicates that much of the flow across the fans is channelized rather than sheet-like, and Menard (1955, p.253) suggests that large flows would tend to split into two parts - the dense bottom layer of coarse sediment following the channel, while the upper layer overtops the levees and spreads radially as a sheet-like flow to deposit finer material over the surrounding areas. This kind of behaviour agrees closely with that deduced by the present author from the sedimentary features of the Singing Creek Formation. Hand and Emery (1964) also emphasize the importance of channelized flow, with the main part of the current being "trimmed" to fit the channel and the excess fine material being lost laterally as the levees are overtopped.

As details of the modern fans became available, several ancient flysch deposits were interpreted as being of this type. The Miocene "Tarzana fan" described by Sullwold (1960) is a particularly interesting example, since it occurs in the Los Angeles basin, a Tertiary equivalent of the present-day off-shore basins discussed above. The fan deposit is characterized by lenticular sand bodies consisting of graded and non-graded beds, in a matrix of clastic shales with interbedded thin sandstones and cherts. This situation closely reflects that of the Singing Creek Formation. In a later review paper, Sullwold (1961) emphasizes the distinction between basin-floor turbidites, which tend to be blanket-like, extensive and fine-grained, and submarine fan turbidites, which may be either channel-like or blanket-like and tend to be laterally variable and coarser grained.

Walker (1966a), in an important paper, interpreted the Carboniferous Shale Grit and Grindslow Shales of northern England as being mainly submarine fan deposits. This sequence has many features in common with the Singing Creek Formation, e.g. alternating sandstone and siltstone zones; graded, non-graded and composite bedding in the sandstones; sequences up to 100 feet thick composed almost wholly of thick-bedded, non-graded sandstones; all gradations from thin sandy laminae through thin graded sandstones to thick, non-graded sandstones; marked lateral variability such that sections more than a mile or so apart could not be correlated. Parts of a number of deep channels filled with typical graded and non-graded sandstones are exposed within the sequence, and were described in detail in a separate paper (Walker, 1966b). The channels are up to at least 70 feet deep, but their total width is doubtful since only one side or edge of the channel can be seen in most cases. Some of the channels show evidence of lateral migration during up-building. One example shows the massive sandstone in the channel passing laterally into normal graded sandstones beside the channel. These graded sandstones, and the associated siltstones which occur beside the channel, are interpreted as levee deposits. Walker argues convincingly that the channels were formed by turbidity currents rather than by slumping or by other kinds of ocean currents.

Two other recent interpretations of ancient deposits in terms of submarine fans are those of Jacka and St. Germain (1967, abstract only), and Harrison and Jacka (1967, abstract only). It is probable that many of those previously-described flysch-type sequences which contain conglomerates ("fluxoturbidites"), numerous composite sandstones, thick non-graded sandstones etc. are also of this type (e.g. Natland and Kuenen, 1951; Wood and Smith, 1959; Basset and Walton, 1960; Marschalko, 1961, 1964; Stanley, 1963; Unrug, 1963; Dott, 1963; etc.).

Comment on Walker's proximality index

Walker (1967) attempts to relate different kinds of turbidites to proximal and distal environments (with respect to source), and develops a "proximality index" based on the sequence of internal

structures shown by each bed. "Proximal turbidites" are thought to be characterized by thick bedding, preservation of the basal structureless division (a) in most beds, coarse grainsize, frequent composite bedding, variability of bed thickness, poorly developed grading, etc. "Distal turbidites", on the other hand, are thought to be characterized by thin bedding, fine grainsize, lack of division a in many beds, regular parallel bedding, well-developed grading, abundant lamination and ripples, high proportion of mudstone, etc. Although the obvious analogy of proximal turbidites with submarine fan deposits, and of distal turbidites with basin-floor deposits, is not actually made by Walker, this is certainly implied in the use of the terms proximal and distal.

The use of the "proximity index", which is a measure of the proportion of beds beginning with each Bouma division (a, b or c), as an indicator of nearness to source is unlikely to be effective in submarine fan deposits because it ignores the lateral variability of these deposits. It seems fairly well established that within submarine fans there is likely to be a fairly marked differentiation into lenses or channel-like units consisting of sand and gravel beds (i.e. "proximal turbidites"), and inter-channel blanket sediments consisting mainly of laminated siltstones and fine sandstones (i.e. "distal turbidites"). The two types are laterally equivalent and are deposited at the same distance from the source. Measurement of the "proximity index" through any vertical section of such deposits will tend to show alternating zones of proximal and distal turbidites, requiring interpretation in terms of oscillating source position. Consideration of lateral variability would probably help to explain some of the anomalies which have arisen using this technique (Walker, 1967; Walker and Sutton, 1967).

3. Conclusions

The Singing Creek Formation is interpreted as a submarine fan complex formed at the base of a steep submarine slope. The position of this slope was probably controlled by the fault system around the western and southern flanks of the range (Pokana Fault, Reeds Creek Fault etc.). Sediment flow down the slope may have been largely via submarine canyons. Gravel, sand and finer sediment from

the (narrow ?) shelf area were carried by density underflows, of the type described previously, and deposited on the fans as the flows decelerated. Earthquake activity along the fault line may have triggered many of the slides which developed into fast-moving underflows.

The sandstone-conglomerate zones are interpreted as being mainly channel deposits or depression-fillings, wherein most of the deposition was from the dense basal and frontal parts of the flows. The intervening siltstone zones are considered to represent blanket deposits from the marginal and rearward parts of the flows as they overtopped the channels and spread laterally. The sequences of interbedded sandstone and siltstone are possibly levee deposits or near-channel deposits. Some of the silt may have been derived from low-density, low-velocity turbidity currents as suggested by Moore (in Buffington et al., 1967).

CHAPTER 3

SEDIMENTOLOGY OF THE GREAT DOME SANDSTONE

A. INTRODUCTION

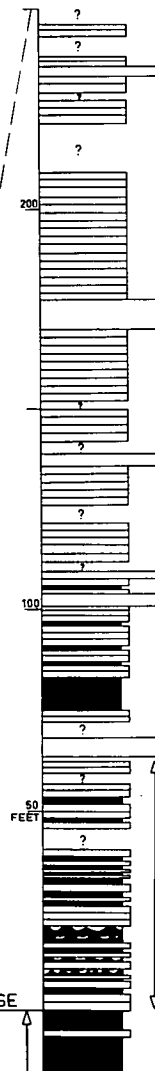
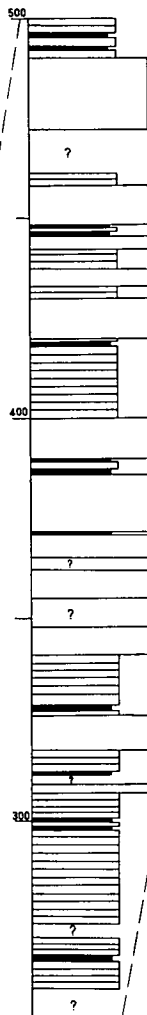
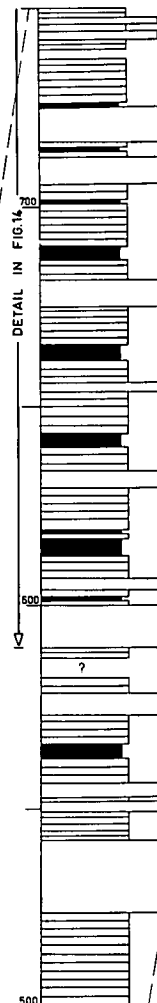
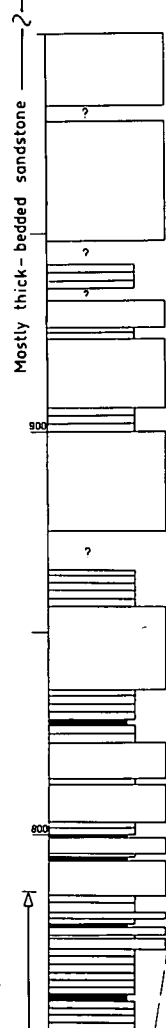
The Great Dome Sandstone is about 1700 feet (510 metres) thick in the type area at Staircase Rocks, where it conformably overlies the Singing Creek Siltstone and forms the upper part of the Denison Group. The formation becomes coarser-grained upwards and passes transitionally into the Reeds Conglomerate. Sparse fossil evidence suggests it is predominantly of Upper Cambrian age. The present study of the sedimentology is necessarily limited because of space and time considerations, and there has unfortunately not been time to treat it as fully as the Singing Creek Formation. A great deal more detailed work could be done, since the formation is well exposed and shows a wealth of sedimentary structures reflecting a variety of shallow water environments.

The formation is petrologically fairly simple, consisting mainly of quartzose sandstones (quartz arenites of Dott, 1964, and Gilbert, 1954) containing varying amounts of mica (mostly muscovite), and micaceous siltstones containing varying amounts of quartz. The abundance of quartzite and quartz-schist rock fragments and pebbles, and the predominantly easterly-directed palaeocurrents, indicate that the Precambrian rocks of the Tyennan Geanticline were the source sediments for most or all of the formation.

The formation can be considered as consisting of two major rock types, viz. thick-bedded, coarse-grained to conglomeratic sandstone, showing large-scale cross-bedding, and thin-bedded or flaggy sandstone with minor siltstone, which alternate in a cyclic sort of pattern throughout much of the formation (Fig. 13). Dark micaceous siltstone constitutes a third but relatively minor rock type, forming beds up to five feet thick. The thick-bedded sandstone units are subordinate to the thin-bedded rocks in the lower half of the formation, herein informally called the Cyclic Facies, but predominate almost to the exclusion of the thin-bedded rocks in the upper part. A number of lithofacies can be identified in the thin-bedded rocks, and in at least some

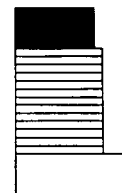
FIGURE 13

TOTAL THICKNESS 1700 feet



SECTION THROUGH CYCLIC FACIES OF GREAT DOME SANDSTONE

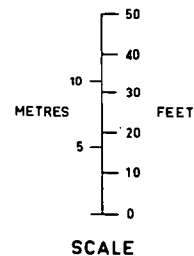
KEY



Siltstone predominant.

Thin-bedded sandstone & minor siltstone; laminated & cross-laminated; abundant worm burrows & bioturbation; also primary current lineation, ripple marks. "flaser" bedding, rare gastropods & inarticulates; low-angle cross-bedding.

Thick-bedded sandstone, conglomeratic in places; large-scale cross-bedding; erosional base.

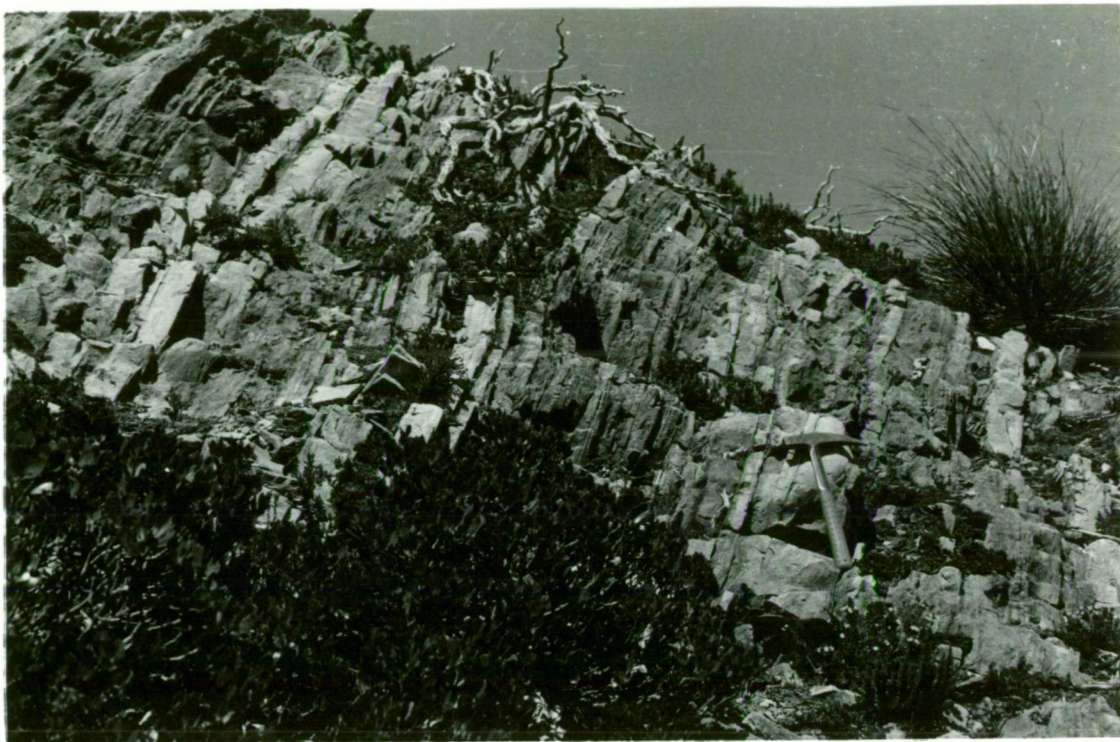


parts of the formation these appear to be arranged in a cyclic pattern. Much of the following discussion concerns the interpretation of the lithofacies and the origin of the cycles.

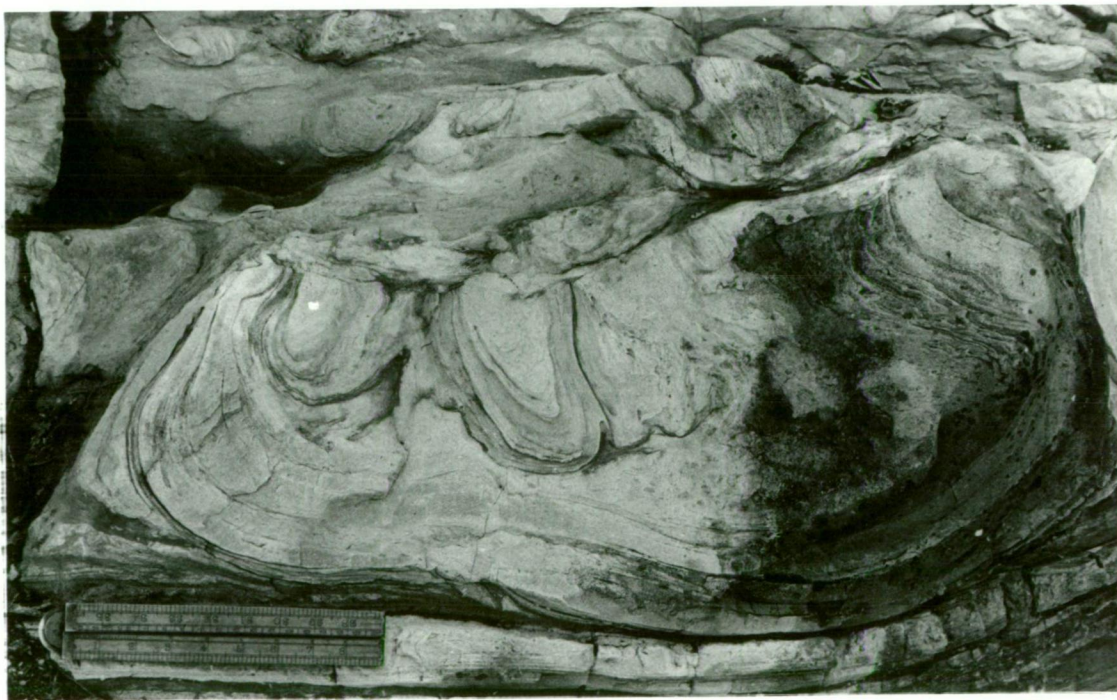
B. BASAL TRANSITION BEDS

The base of the formation is transitional downwards into siltstones forming the top of the Singing Creek Siltstone (Fig. 13). At Staircase Rocks about 200 feet of these siltstones lie between the basal sandstones of the Great Dome Formation and the first major sandstone zone of the Singing Creek Formation. The Transition Beds comprise 60-100 feet of interbedded dark siltstones and light-coloured quartzose fine sandstones (Plate 32A), with the proportion of sandstone increasing upwards. The sequence is characterized by ubiquitous flat lamination (without parting lineation for the most part) and by interbedding of sandstone and siltstone on all scales from fine laminae up to beds 10 inches (25 cm) thick. Flow-rolls, or pseudonodule-type deformational structures, are also characteristic of the sequence, and range in size from an inch across to more than three feet wide. They include isolated, clustered (ball-and-pillow) and composite types (Plate 32B), and appear to have been formed by the foundering of a sand layer into either silt or water-saturated sand. In many cases the foundering has occurred only over a restricted area, and the sandstone layer is elsewhere undisturbed.

Other features typical of the Transition Beds include low-angle erosion surfaces (shallow channels?), small sand-filled or silt-filled channel structures (some of which show syn-depositional load deformation and resemble flow-rolls), worm burrows, biogenic reworking, "quicksand" structures, and apparent slump structures. In the upper part, current structures such as ripple marks, cross-lamination, parting lineation etc. begin to appear, but these are rare in the lower beds. The sequence is tentatively interpreted as being mainly sub-tidal facies deposited at the seaward edge of a delta or estuarine complex at sufficient depths to avoid strong reworking by waves or tidal currents.



A. Outcrop of Transition Beds at base of Great Dome Sandstone, Staircase Rocks. Interbedded dark micaceous siltstone and light-coloured fine sandstone.



B. Large composite flow-roll, Transition Beds. Surrounding material mainly structureless sand with remnants of laminated sand. Internal structures apparently formed by sand extrusion from centre of main flow roll. Dark silty bands may have formed water-seals.

C. CYCLIC FACIES

The lower 900 feet of the formation shows a distinct tendency towards a cyclic or alternating pattern of lithofacies (Fig. 13). This is mainly evident as an alternation of thick-bedded coarse sandstone units with zones of thin-bedded sandstone and siltstone (Plate 33A, B). The thick-bedded units crop out strongly to form a striking series of sub-parallel shelves or steps along the steep western slopes of the range, while the thin-bedded zones form relatively negative areas and are less well exposed. The thick-bedded units are mainly 5 to 20 feet (1.5-6 m) thick, while the thin-bedded zones range up to more than 50 feet (15 m) thick. Detailed examination of the well-exposed thin-bedded zones generally reveals several distinct lithofacies defined by rock types, sedimentary structures, and to a lesser extent the faunal content. In some sections at least, these lithofacies occur in a more or less constant sequence with respect to the thick-bedded units, forming a cyclic pattern involving 5 or 6 members.

Fig. 14 shows a detailed section through 200 feet (60 m) of the formation (from 590 to 790 feet on Fig. 13) as exposed on two adjacent small spurs at Staircase Rocks. This section is the best-exposed part of the formation, and by correlating and tracing units from one spur to the other (a strike distance of about 250 feet or 75 m) an almost complete section through a number of cycles was obtained. The following account is based mainly on this detailed section, which appears to be fairly representative of the Cyclic Facies in most areas, although this has not been confirmed in detail. A legend and interpretation for Fig. 14 is given in Fig. 15.

1. Definition of Cycles

The number of cycles considered to be present depends to a large extent of what is regarded as constituting a cycle. If each alternation of a thick-bedded unit and a thin-bedded zone is regarded as a cycle, then at least 12 are present within the 200 foot interval of Fig. 14. Some of the thin-bedded zones, however, are only two feet or less in thickness, and some of these are merely discontinuous lenses sandwiched between two thick-bedded units which amalgamate laterally to



A. View of Cyclic Facies of Great Dome Sandstone in type area at Staircase Rocks. Thick-bedded sandstone units form prominent shelves separated by zones of thin-bedded rocks. Flagstone Knoll in background.

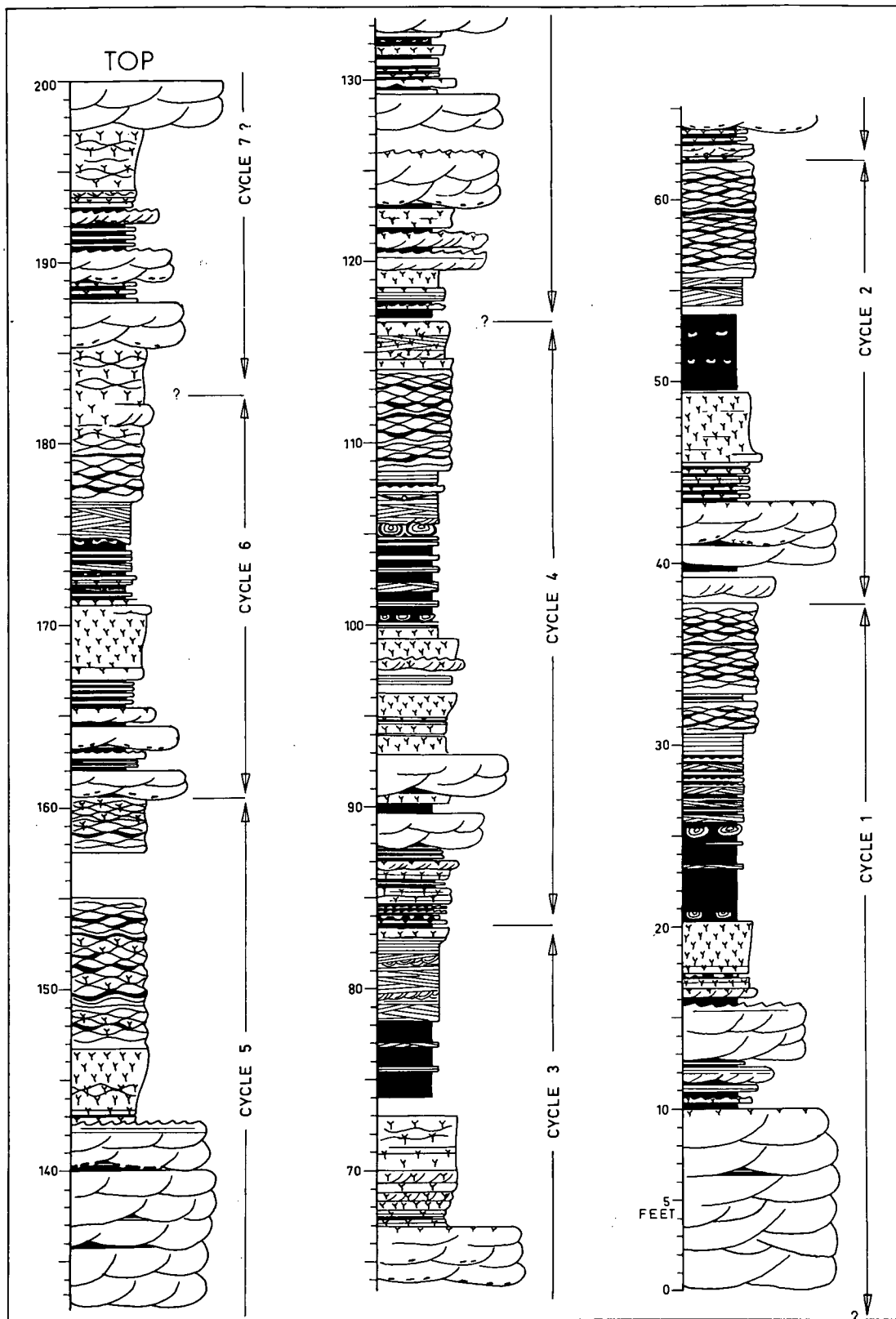


B. Thick-bedded unit of Lithofacies A disconformably overlying thin-bedded zone. Note large-scale cross-bedding and the group of smaller thick-bedded units above the main one.

form a single composite unit. Most of the thicker thick-bedded units appear to be composites formed by amalgamation of smaller units in this way, and in general there is a strong tendency for the thick-bedded units to occur as "clusters" separated by thin intercalations of thin-bedded material (e.g. Plate 33B). These clusters of thick-bedded units can therefore probably be regarded as constituting one member of the cycle rather than several small cycles. The major thin-bedded zones are mostly at least 10 feet thick and characteristically show several lithofacies arranged in a more or less constant sequence. Each such zone, together with the associated thick-bedded unit or cluster of such units, is herein considered as representing a cycle. Six such cycles are represented in Fig. 14.

The validity and significance of so-called cyclic sediments have recently been discussed in detail by Duff et al. (1967). These authors emphasize the variability of most cyclic deposits and the need for objective treatment, and recommend that distinction be made between the most commonly-occurring or modal cycle and the complete cycle or composite sequence. In the present case there is probably not sufficient data on a sufficient number of cycles to determine the modal cycle, and most of the following discussion is concerned with interpreting the origin of the various lithofacies and of what appears to be the composite sequence.

In order to describe and interpret the sequence it is necessary to make some sort of grouping of rock units into lithofacies, since it is obviously impossible to describe each bed or lamina individually. Using rock type and sedimentary structures it is possible to subdivide the sequence into five main lithofacies which recur in most cycles and which appear to have genetic significance. These are: (A) thick-bedded coarse sandstone units; (B) thin-bedded sandstone and siltstone associated with the thick-bedded units; (C) siltstone; (D) sandstone with low-angle cross-bedding; (E) cross-laminated sandstone with marine fossils. Lithofacies A and B are usually interbedded. A complete cycle, or composite sequence, comprises ABCDE. Cycles 1, 2, 4, and 6 of Fig. 14 are of this type, while Cycle 3 is ABCDA and Cycle 5 is ABE.



DETAILED COLUMNAR SECTION THROUGH PART OF CYCLIC FACIES
OF GREAT DOME SANDSTONE AT STAIRCASE ROCKS
SHOWS SIX COMPLETE CYCLES
FOR EXPLANATION SEE FIG. 15

FIGURE 14

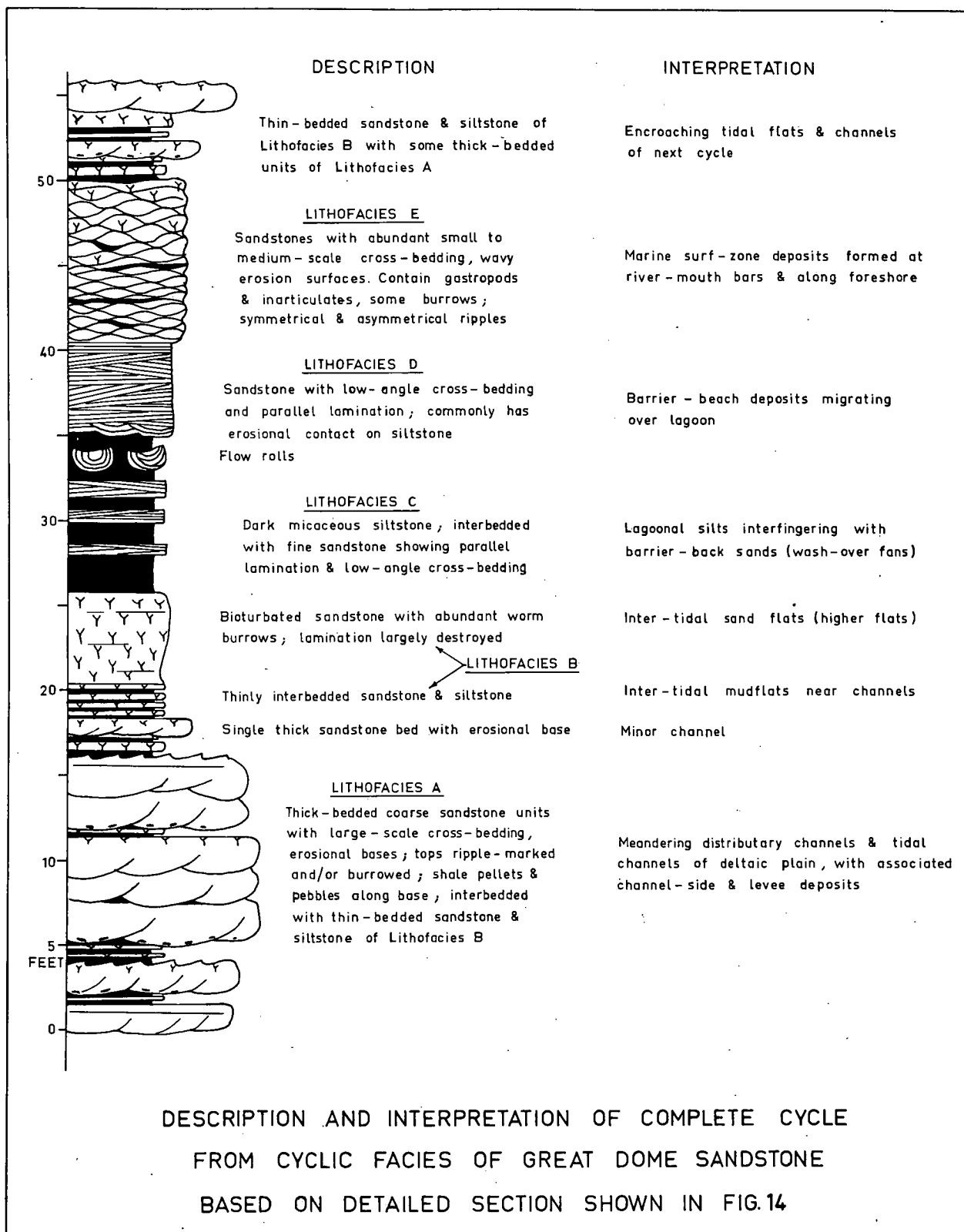


FIGURE 15

2. Lithofacies A. Thick-bedded Coarse Sandstone Units

These units range in thickness from about 2 feet (0.6 m) to more than 30 feet (9 m), although most of the thicker units appear to be composites. They consist of light-coloured (white to pale grey) coarse-grained to conglomeratic, pure quartz sandstone (mature quartz arenite of Dott, 1964), and characteristically show large-scale cross-bedding throughout (Plate 33B). The base of each unit is usually abrupt and erosional on the underlying thin-bedded rocks, the erosion surface commonly showing broad trough-like undulations (maximum relief about 18 inches) similar to those forming the large-scale cross-bedding. Pebbles and intraformational shale pellets, and rare gastropods probably derived from reworking of the underlying beds, occur along the surface in places, and large load structures may be present where the base rests on a siltstone layer. The upper contacts of the thick-bedded units are usually fairly sharp and show ripple marks and/or worm burrows, but gradational contacts into thin-bedded sandstone also occur.

The cross-bedding is predominantly of the large-scale trough type (Pi cross-stratification of Allen, 1963), but planar sets (Omikron-type of Allen) and sets which appear to be intermediate between the two types (i.e. with gently undulating lower bounding surface) are also present. Individual sets are usually 1-3 feet thick, and the number of sets within a unit ranges from one up to 20 or more. The troughs range in width from about 12 inches (30 cm) to about 20 feet (6 m), the majority being more than 3 feet wide. Foresets are tangential to the base of the set in most cases. Preliminary observations indicate that the cross-bedding in most units is uni-directional towards the east, and no examples of reversed or chevron cross-bedding have been seen. Further detailed analysis of cross-bedding directions within these units is required, however.

The majority of the units maintain a fairly uniform thickness across the outcrop width (200-400 feet or 60-120 metres), and some of the thicker units appear to continue along strike for at least half a mile. Some of the thinner units, however, consist of only one or two beds, and show considerable thickness changes over distances of a few hundred feet. This is in part due to the variable erosion at the base

of the unit, but there are a few examples where a thick-bedded unit disappears laterally into thin-bedded sandstone.

Interpretation

Similar units of cross-bedded sandstone resting on an erosion surface have been described from ancient and modern deposits by many authors, and are now confidently interpreted as being formed by a meandering stream channel of some kind (i.e. fluvial, deltaic, or tidal). Much of the original work concerned modern river deposits (e.g. Happ et al., 1940; Jahns, 1947; Sundborg, 1956; Wolman and Leopold, 1957), but the synthesis of this work and the recognition of these "fining-upwards sequences" in ancient deposits must be credited mainly to J.R.L. Allen (1964, 1965a, 1965b). Although most deposits of this kind occur in fluvial sequences, similar units are formed by meandering stream channels in deltaic (e.g. Oomkens, 1967), estuarine (e.g. Oomkens and Terwindt, 1960; Land and Hoyt, 1966) and tidal-flat environments (e.g. Van Straaten, 1954a, 1954b, 1961 etc.; Klein, 1963, 1965; Evans, 1965).

The method of formation of the channel sequence is similar in all cases. The stream migrates laterally across the floodplain (or estuary or tidal flat) by erosion of the outer (concave) bank of each meander, producing a sub-horizontal erosion surface at the level of the bottom of the channel. This lateral erosion is balanced by deposition of sand within the channel and on the inner bank or point bar, so that the result of the migration is a sheet of channel and point-bar sands lying disconformably on the underlying finer-grained floodplain sediments. The process is known as lateral accretion. Coarse sand and gravel, as well as mud flakes and shell fragments, may be concentrated as lag deposits at the base of the channel.

Most of the deposition within the channel involves bed-load material and is accomplished by migrating megaripples (also called dunes, sand waves, bars). These migrate in the direction of the strongest current, which is usually seaward but may be in both directions or wholly landward in some tidal situations. Large-scale cross-bedding is produced by these megaripples, the form depending on the shape of the megaripples and of the associated scour features

(e.g. Frazier and Osanik, 1961; Harms et al., 1962; Bernard and Major, 1963; Lane, 1963; Harms and Fahnestock, 1965). Trough-type is recorded as being the most common. On the higher parts of the point bars there is usually deposition of suspended material during high flows as well as bed-load material, and these deposits tend to be finer-grained and more thinly-bedded, producing the fining-upwards sequence.

The stratigraphy of the Great Dome Formation also suggests that the thick-bedded units represent stream channel deposits. The units become thicker, coarser-grained and more predominant upwards (Fig. 13), and at the top there is a transitional passage into the Reeds Conglomerate. The latter consists mainly of red conglomerates and sandstones lacking any marine fauna, and is interpreted as an alluvial fan deposit. The cross-bedded conglomeratic sandstones and fine conglomerates which form the bulk of the upper part of the Great Dome Sandstone can thus confidently be interpreted as stream-channel deposits, and the obvious similarity to the thick-bedded units in the Cyclic Facies suggests these are also channel deposits.

Throughout most of the formation the rocks interbedded with the thick-bedded units contain abundant worm burrows, and there are also marine gastropods and inarticulates in the Cyclic Facies. Thus the channel deposits must have been formed in a paralic environment rather than under truly non-marine conditions. Delta distributaries, meandering estuaries, or tidal channels are the main possibilities, but distinction between these is difficult on present knowledge. The occurrence of worm burrows within the channels in a few places, the intimate association with thin-bedded rocks containing abundant worm burrows, and the presence of (reworked ?) gastropods in the bottoms of some units, indicates that at least some of the channels were subject to tidal influence. However, the apparent predominance of uni-directional east-dipping (i.e. seaward) cross-bedding and the lack of observed cases of reversed cross-bedding strongly suggests that river currents rather than tidal currents were mainly responsible for deposition of the units. Some of the smaller units, however, particularly those which are not persistent laterally, may have been formed by tidal creeks associated with the major distributaries.

The tendency for the thick-bedded units to occur as clusters of several units interbedded with flaggy sandstone and siltstone indicates that the distributary system remained over an area for a sufficient time for a sequence of channel deposits to build up. Successive channels must have migrated over one another as the area subsided. The channels were not all of the same size, since the clusters generally show one or two major units and several minor ones.

3. Lithofacies B. Thin-Bedded Sandstone and Siltstone with Abundant Worm Burrows

Interbedded with and overlying the thick-bedded units, and generally underlying them to some extent also, are thinly-interbedded quartz sandstones and micaceous siltstones showing parallel lamination, ripple marks, cross-lamination, irregular bedding contacts, and abundant worm burrows. Parting lineation, and in a few cases current crescents around worm tubes, are developed within some of the flat-laminated sandstone horizons. In some places this thinly-bedded material fills broad shallow depressions on the tops of the thick-bedded units. Flow-roll structures consisting of laminated sandstone surrounded and intruded by siltstone occur on some horizons (Plate 34A). In the example illustrated, the siltstone appears to have intruded at a number of points, so that many of the rolled-up structures remain connected and have widely-varying orientations. Structures resembling raindrop imprints also occur within this lithofacies in a few places. These consist of numerous small circular depressions 2-5 mm across almost covering the bedding plane. In one case they occur as moulds on the underside of a sandstone layer resting on a thin silty band.

The interbedded sandstones and siltstones generally pass upwards into an interval consisting predominantly of sandstone and showing very pronounced bioturbation (Plate 34B). Within this interval, which is usually several feet thick, the original lamination has been almost completely destroyed by burrowing, leaving only remnants of parallel lamination and small- to medium-scale cross-bedding. Some of these bioturbated horizons are reddish-brown in colour due to abundant limonite, and some have irregular veins of limonite running through them.



A. Lithofacies B. Flow-rolls of laminated sandstone surrounded by dark sandy siltstone. Overlain and underlain by bioturbated silty sandstone. Note burrowed top of cross-bedded clean sandstone unit (minor channel ?) beneath rule.



B. Bioturbated sandstone of upper part of Lithofacies B (inter-tidal sand flats ?). Note burrowed remnants of flat-laminated sandstone. Outcrop is heavily limonite-stained.

This feature probably indicates some degree of subaerial exposure. The burrows are predominantly of the U-shaped variety (cf. Corophioides, Hantzschel, 1962), but also include bedding-plane burrows and massed vertical tubes. The tops of many sandstone layers show a characteristic "birds-feet structure" consisting of numerous overlapping depressions left by the bottoms of U-shaped burrows in the overlying bed. Similar horizons of brown bioturbated sandstone occur just beneath the main thick-bedded channel units in some cycles.

Interpretation

The intimate association of this lithofacies with the thick-bedded units of Lithofacies A suggests that the rocks were deposited adjacent to and perhaps partly within the channels in which the cross-bedded coarse sandstones accumulated. The small-scale interbedding of sandstone and siltstone suggests variable currents, with periods of slack water, as might be expected on the higher parts of the channel sides. Allen (1965c, pp.569-571) describes similar thin-bedded sands and silts from channel-sides and levees along meandering distributaries towards the seaward edge of the Niger delta. These sediments overlie the coarse sands deposited in the channels and on the lower parts of the point bars. The levee deposits are only intermittently submerged, and show abundant animal burrows, giving rise to "pronounced mottles bordered by red-brown oxidized rims". Small beaches developed along some of the channel margins produce flat-laminated sand with primary current lineation.

The sediments of Lithofacies B also appear to be closely analogous with modern tidal-flat sequences as described by various authors, particularly van Straaten (1954a, 1954b, 1961), Evans (1965) and Reineck (1964). According to these authors, the channels which occur in the lower part of the tidal flats are bordered by "mud flats", characterized by inter-laminated sand and mud, and these are followed at a slightly higher level by sand flats ("Arenicola sand flats" of Evans) in which burrowing by worms is sufficiently intensive to destroy the lamination in places. Such a sequence is characteristic of Lithofacies B. Some evidence of subaerial exposure can be found on most tidal flats, and in Lithofacies B this is shown by the raindrop imprints and by the

abundant iron oxide in some of the bioturbated horizons. Surprisingly, perhaps, there appear to be few if any mudcracks preserved in these rocks.

Lithofacies B is therefore tentatively interpreted as representing the channel-sides, levees and tidal flat areas bordering the meandering distributary channels in which the cross-bedded coarse sands were deposited (Fig. 15).

4. Lithofacies C. Siltstone

Overlying the bioturbated sandstone at the top of Lithofacies B in most cycles is a variable thickness (up to 5 feet) of dark grey micaceous sandy siltstone, usually interbedded and finely inter-laminated with fine sandstone. The contact with Lithofacies B is usually sharp but not erosional. The siltstone is missing in Cycle 5 and is subordinate to sandstone in Cycle 6. The proportion of laminated sandstone increases upwards, forming a transitional contact with the Lithofacies D. The interbedding of these two rock types indicates that the two lithofacies represent closely-related environments.

The siltstone generally shows only vague parallel lamination or slightly wavy lamination, whereas the interbedded sandstones show well-developed parallel lamination (Plate 35A) and in some cases low-angle cross-bedding. Worm burrows and mottling occur in some sections, the burrows being filled with sand in many cases (Plate 35A). Small-scale deformational structures apparently formed by extrusion of mud through a sandy layer occur on some horizons (Plates 35A, 36A), while larger-scale structures (flow-rolls or pseudonodules) are characteristically developed at sandstone-siltstone contacts in the higher levels, as described later.

In thin section the siltstone is rather poorly sorted and consists mainly of quartz grains with 30-50% micaceous minerals, minor carbonate, and accessory pyrite, tourmaline and zircon. The quartz grains range from fine sand to medium silt grade, and are mostly angular. The mica is mainly muscovite and sericite with lesser biotite and chlorite. In some patches the micaceous minerals have an irregular box-work appearance, but elsewhere are predominantly parallel to bedding. The fine sandstones are similar in composition, except that quartz grains



A. Dark siltstone of Lithofacies C, with intercalations of laminated fine sandstone. Note sand-filled worm burrows; deformation structure; limonite pits in sandstone. Scale in inches. From Cycle 6 of Fig. 14.



B. Rafts and flow-rolls of laminated Lithofacies D sandstone surrounded by sandy siltstone of Lithofacies C. Scale at top in inches. From Cycle 1 of Fig. 14.

are more predominant, suggesting that the sands may have been derived from slight reworking of the silty material.

Interpretation

The predominance of siltstone suggests quieter conditions than for the other lithofacies, with much of the deposition being from suspension. The interbedded sandstones show low-angle cross-bedding and other structures suggestive of beach deposits, as discussed below, and the silts appear to have been deposited in a low-energy lagoonal or bay environment adjacent to small beaches. Further consideration of this environment is incorporated in the discussion of the next lithofacies.

5. Lithofacies D. Sandstone with Low-Angle Cross-Bedding

Interbedded with and overlying the siltstones of Lithofacies C are fine-grained light-coloured quartzose sandstones characterized by fine parallel lamination and low-angle cross-bedding (Plates 35, 36). Sandstones of this type may first appear near the base of the siltstone horizon, but they become more predominant upwards until the siltstone disappears. The top of the lithofacies generally consists of a slightly coarser, siltstone-free sandstone unit, 2-4 feet thick, with an erosional base (Plate 36A). Contacts between sandstone and siltstone layers are mostly gradational in the lower part of the unit (Plate 35A), but tend to be sharper and erosional in the upper part.

In thin section the sandstones consist of three main components, viz. quartz grains (60-70%), muscovite (10-20%) and carbonate cement (10-20%), with minor limonite and glauconite and accessory tourmaline and zircon. Weathered samples are slightly limonite-stained, and contain numerous small rounded blebs of interstitial limonite as well as disseminated limonite derived from weathered carbonate. The rounded blebs of limonite, which are up to 0.5 mm diameter, appear as pits on many weathered surfaces (e.g. Plates 35, 36) and may be a primary feature (gas pits?). The quartz grains are mainly fine to very fine sand grade, but include medium and coarse sand in the uppermost unit. The grains are mainly angular to sub-angular, and many show undulose extinction. The micas are strongly aligned parallel to bedding, and in

some cases are concentrated into thin laminae. Lamination planes usually show abundant mica except in the coarser-grained upper unit, where mica is much less common.

Structures

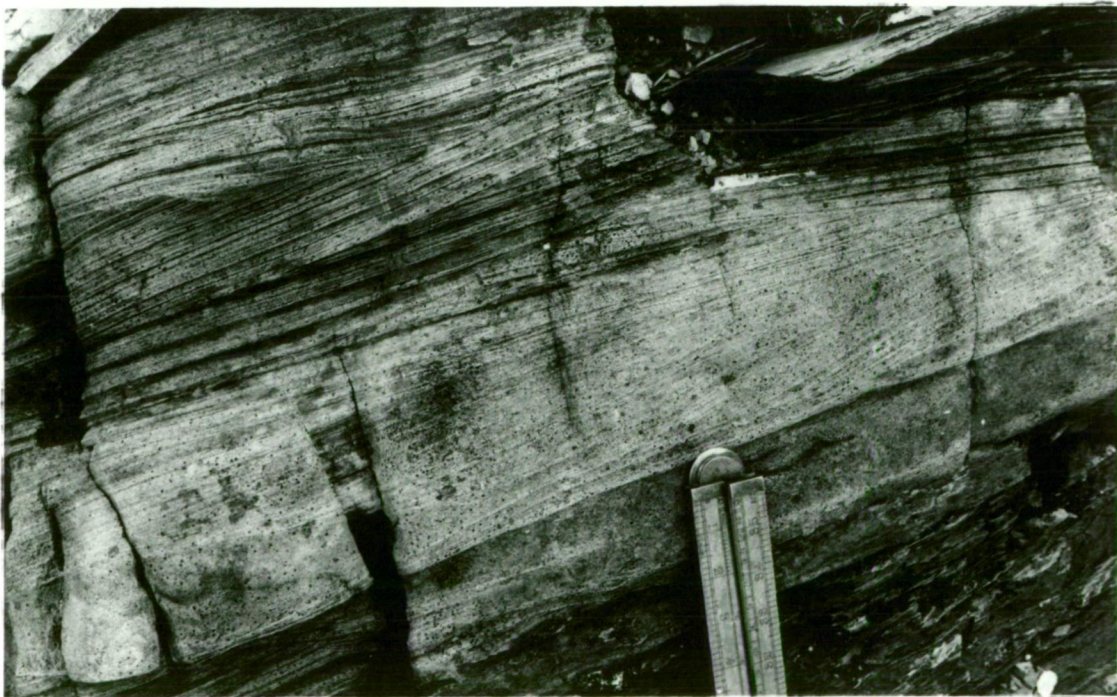
Parallel lamination and low-angle cross-bedding are characteristic of this lithofacies, and the latter structure distinguishes these rocks from all others in the sequence. Other structures include small- and medium-scale cross-lamination, ripple marks, worm burrows, scour- and-fill structures, pseudonodules (flow-rolls), and parting lineation. The low-angle cross-bedding actually consists of a series of interfering low-angle erosion surfaces, usually spaced only a few inches apart, most of which are overlain by laminae which are parallel to the surface. In some cases, however, the laminae are tangential to the surface (e.g. lower part of Plate 36B). Some of the surfaces truncate thin bands of siltstone, and in a few cases shale pellets are preserved along the surface. Wedging-out of thin sandstone and siltstone layers by erosion is a characteristic feature.

The angle between successive surfaces ranges from one degree or less to about 15° , with larger angles at the sides of some local scour features. The very low angle surfaces are difficult to recognize except in good outcrops, where they appear to be fairly abundant. It seems that deposition of a great many of the laminae was preceded by slight erosion at one place or another, the amount of erosion determining the prominence of the surface. Some surfaces become indistinguishable away from the area of maximum erosion. The form of the surfaces in three dimensions is difficult to reconstruct, since they are seen only in section. Some appear to be planar and fairly extensive, while others are distinctly trough-shaped with the infilling laminae becoming flatter upwards and tending to smooth off the depression (Plate 36A). Parting lineation is associated with the parallel laminae and low-angle cross-laminae in the upper levels of this lithofacies in a few places, but is by no means a general feature.

Small- to medium-scale cross-lamination is locally developed but is seldom a dominating feature. Sets are mostly of the trough type, with a relief of up to 8 inches (20 cm). The infilling laminae are



A. Erosional base of main upper sandstone unit of Lithofacies D in Cycle 6. Note truncated horizon of deformed sandstone; trough-like feature with low-angle cross-bedding.



B. Low-angle cross-bedding in Lithofacies D sandstone of Cycle 2. Note numerous erosion surfaces, also limonite pits. Scale in Inches.

usually tangential to the sides, such that the troughs become smoothed off upwards. The structures resemble smaller-scale versions of the trough-like features described above, and appear to be related to local scours rather than to ripple migration. Some tabular sets are also developed, in which the dip of successive foresets gradually decreases downcurrent until the laminae are sub-horizontal (e.g. bed near top of Plate 35B). Here again, the structure appears to have resulted from aggradational smoothing-off of a bottom irregularity. Definite small-scale ripple marks are preserved in a few places, however. Some of these have internal foresets as in normal current ripples, but in a few cases the ripples are symmetrical and appear to be erosional features developed on flat-laminated sand - this type may be wave-produced.

Flow-rolls or pseudonodules are characteristically developed in the horizons of interbedded sandstone and siltstone. They range in size from small ball-shaped structures an inch or so across to broad flat rafts up to 10 feet wide (Plate 35B), and consist of laminated sandstone of Lithofacies D surrounded and intruded by sandy siltstone of Lithofacies C. The siltstone commonly shows a weak cleavage wrapping around the flow-rolls. Most of the smaller structures have upturned edges and rounded outlines, but some of the larger rafts have sharp, truncated edges with only slight up-turning of laminae. Some examples show a series of small faults along which the edges of the flow-roll have been displaced progressively upwards. Some thin sandstone layers show incipient flow-roll development along some sections where "flames" of silt have penetrated only part-way through the bed. Possibly the intrusive pressure was released when the bed was ruptured at another point. In other cases, undisturbed sections of the bed alternate with sections showing flow-roll development. The upturned edges of the flow-rolls on some horizons have been truncated before deposition of the overlying laminated sandstone, and in some cases it is apparent that upwelling of the silty material has continued after this initial truncation, such that the truncating laminae are themselves disrupted.

The flow-rolls may have been formed either by loading, causing the sand to sink into the mud because of excess density, or by release of water pressure from the buried silty layer, causing it to intrude upwards through the sand to the surface, or by a combination of

both. In some cases the active or intrusive material is mainly sand, and is separated from the flow-roll sandstone by a thin layer of very micaceous silt. In these cases, release of water pressure seems the most likely explanation, and the silty layer may have acted as a seal to allow the pressure to build up to such a level that its release caused rupture of the overlying sand (see also Plate 32B). Similar structures have been described by many authors (e.g. Sorauf, 1965; Potter and Pettijohn, 1963; Selley et al., 1963; Pepper et al., 1954; Emery, 1950; Macar, 1948), and appear to be characteristic of, although not restricted to, shallow marine or fluvio-marine sequences. The example described by Emery (1950) involved Quaternary beach and lagoon sediments.

Interpretation of Lithofacies D

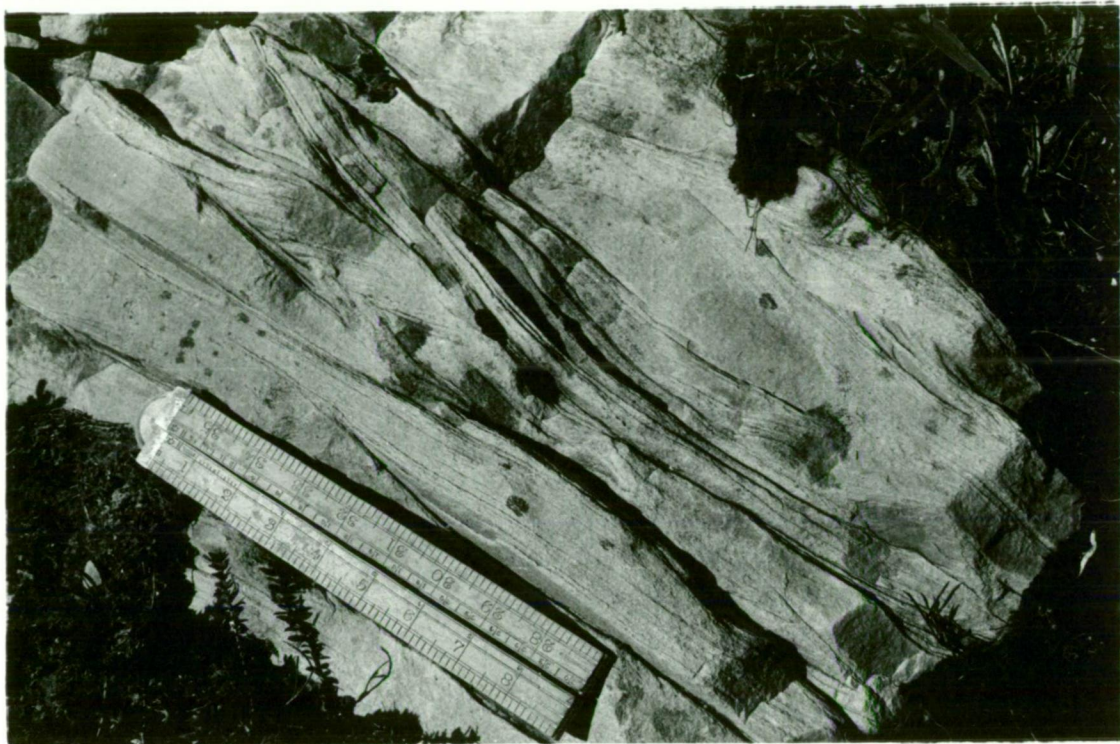
Low-angle cross-bedding of the type shown by Lithofacies D is unusual in sedimentary sequences, and a search of the literature suggests it is characteristic of the beach environment and virtually restricted to this environment. The most comprehensive description of beach structures remains that of Thompson (1937), while others include McKee (1957), Emery (1945), Emery and Stevenson (1950), Otvos (1964), Land et al. (1967), Hoyt (1962), Trefethen and Dow (1960), Soliman (1964), and Hayes et al. (1969, several papers). The cross-bedding develops by aggradation of parallel laminae onto an erosion surface (formed by wave activity, swash, storms etc.) which truncate the underlying laminae at a low angle. The laminae persist for considerable distances parallel to the shore on flat beaches, but tend to be discontinuous and in wedge-shaped or trough-shaped sets on cusped beaches or in sections normal to the shoreline. The nature of the beach sediment depends largely on the wave energy, and high-energy beaches are generally coarse-grained (and may include shingle deposits etc.) whereas low-energy beaches may consist largely of fine sand. Descriptions of very low-energy beaches, such as developed behind barrier islands, are rare in the literature. Other structures typical of beaches include ripple marks (both current- and wave-produced), organic burrows, small to large channels formed by swash or tidal outflow, swash and rill marks, streaming lineation, disturbed lamination due to varying water table, gas pits, high-angle cross-bedding formed by bars or spits etc.

The fine grainsize of the sands and the abundance of mica in the lower levels of this lithofacies suggest a low-energy environment rather than an exposed coastline. As well as this, the interbedding of sandstone and siltstone indicates that the two environments were alternating and transitional for a considerable period of time before sand deposition became predominant. Such conditions might be expected in the marginal areas of lagoons or bays protected by barrier islands, with small low-energy beaches (wash-over fans ?) interfingering with lagoonal muds. Similar interfingering of "bar" sands with "lagoon" silts is described by Johnston and Friedman (1968), who also record flow-roll development at this level. The main upper sandstone unit of Lithofacies D generally has an erosional base and is coarser-grained and less micaceous, possibly suggesting the migration of the main barrier beach over the lagoon. Emery (1950) describes flow-rolls developed in beach sand which had migrated inland onto lagoonal silts in just this manner.

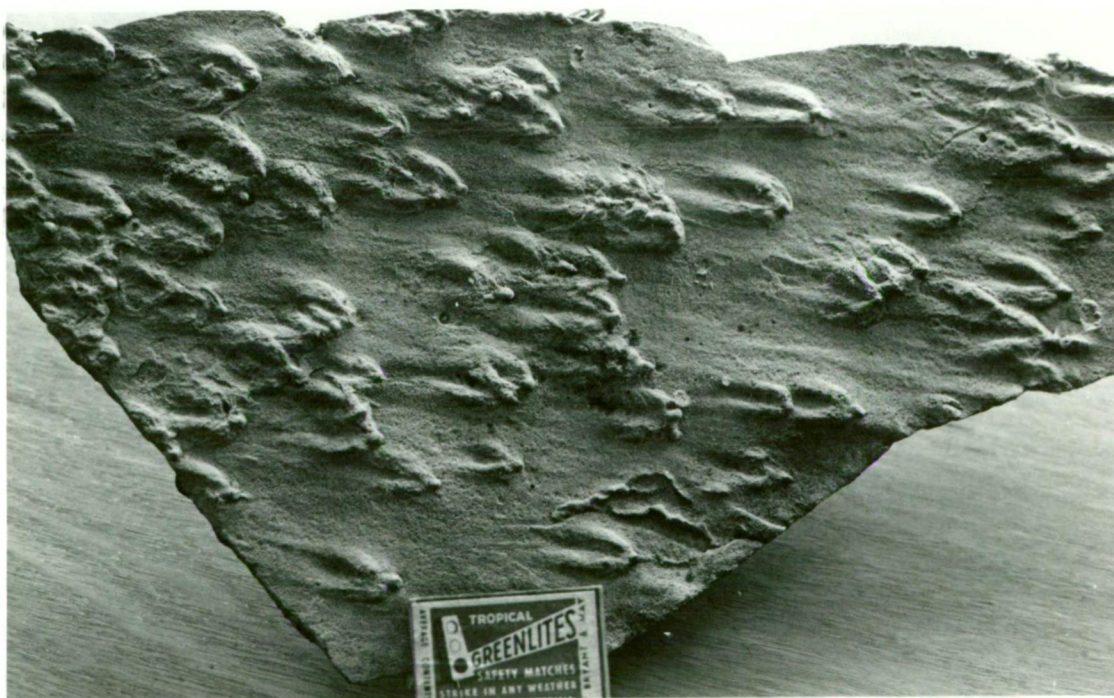
6. Lithofacies E. Cross-Bedded Sandstone with Marine Fossils

Overlying the sandstone with low-angle cross-bedding there is usually a fairly thick sequence of fine to coarse quartzose sandstone showing abundant small- to medium-scale cross-bedding, with some intercalated undulatory bedding and flat lamination (Plate 37A). Thin bands and lenses of siltstone are interbedded with the sandstones in places, forming crude flaser structure. Intergradations from flat lamination to undulatory lamination to cross-lamination are common. Gastropods (cf. *Kobayashiella*) and small, poorly-preserved inarticulates are common at some levels in these sandstones. Worm burrows are also common, although bioturbation is not as intense as in Lithofacies B. The sandstones are usually light-coloured, but some outcrops have a faint pink colour due to limonite-staining.

The cross-bedding is almost wholly of the interfering trough type, with troughs up to about 2 feet wide and 8 inches deep. The foresets in many cases pass laterally into horizontal laminae. The cross-bedded sets are mostly deeply truncated by the overlying set or by wavy erosion surfaces, suggesting considerable reworking of the sediment rather than continuous aggradation. Fully-preserved ripple marks also occur, however, and include symmetrical types, with laminae dipping in



A. Typical cross-bedded sandstone of Lithofacies E, Cycle 1. Note interfering troughs, horizontal lamination, apparent reversals of current direction.



B. Moulds of current crescents around worm tubes on underside of sandstone bed, Lithofacies E. Broken-off ends of tubes are inclined downcurrent.

both directions, as well as asymmetrical current ripples. The ripples range in size from an inch or so in wavelength to those approaching mega-ripple dimensions (wavelength more than 2 feet). The orientation of the cross-bedding appears to be very variable, although detailed measurements have not yet been made. Examples of apparent reversal of current direction can be seen in most outcrops.

Parting lineation occurs on most of the flat-laminated sandstone horizons, and is more common in this lithofacies than in any of the others. Current crescents formed around projecting worm tubes also occur on some of these horizons (Plate 37B). The crescents are always parallel to the associated parting lineation, thereby confirming that the latter structure is current-produced and is parallel to the current. The crescents are best seen as moulds on the underside of the overlying bed, where they form paired ridges extending downcurrent from the broken-off ends of the sand-filled tubes. Most of the broken-off tubes are inclined as though bent over by the current, although their continuations in the underlying bed are usually sub-vertical. Similar structures occur in Lithofacies B in places.

Interpretation

The combination of symmetrical and asymmetrical ripple marks, multi-directional cross-bedding, flat-laminated sandstone with parting lineation, and abundant erosion surfaces, suggests a fairly high-energy environment with considerable reworking and probable wave activity. The presence of gastropods and inarticulates suggests that conditions may have been more fully marine than in the underlying lithofacies. The presence of siltstone bands and lenses indicates periods of quiet water also, and may reflect tidal conditions as inferred for normal flaser bedding (e.g. Reineck and Wunderlich, 1968).

Conditions similar to those deduced above occur on the river mouth bars which project beyond the barrier beaches at the seaward edge of some deltas. Many such bars occur on the Niger delta (Allen, 1965c), where they are "subject to swell waves breaking over them, to littoral currents resulting from wave-breaking generally, and to swift but reversing tidal flows" (*ibid.*, p.580). The sediments consist mainly of sand showing even lamination and cross-lamination, but interbedded sand

and silt are deposited when the river becomes sand-depleted due to shifting of the main depositional area. Similarly, the distributary mouth bars of the Mississippi delta (Coleman and Gagliano, 1965) are constantly subject to reworking by waves and currents, and are characterized by thin, multi-directional trough cross-lamination in sands and silts, with some preservation of wave and current ripples because of the high sedimentation rate.

A similar environment is the foreshore or surf zone which lies just seaward of the main barrier beach environment. The sediments formed in this zone are poorly known since they can normally only be studied in cores. The surface features include long-shore bars and shallow channels, wave and current ripples, megaripples, tidal pools, biogenic burrows, and small areas of flat beach. Reineck (1964) reports laminated sand, ripple bedding, megaripple bedding, rare flaser bedding, and bioturbation from this environment, while Thompson (1937) records laminae of micaceous sand on the floors of tidal pools being buried by steeply-inclined cross-laminae from sand bars containing shell material. Hayes et al. (1969a) suggest that low-tide terrace and surf-zone deposits, consisting of inter-layered plane beds and festoon cross-beds with random orientations, would overlie beach-face sediments in a transgressive sequence. Lane (1963) has interpreted cross-bedded sandstones lying between beach deposits and marine shale as surf-zone deposits, while van der Linden (1963) has attributed rather similar cross-bedded sandstones with "mega-flaser" structure to deposition on a "wave-built terrace".

Lithofacies E is therefore interpreted as representing surf-zone deposits formed around the seaward margin of a deltaic or estuarine complex at the mouths of distributaries and/or along the barrier beaches between distributaries. The position of the lithofacies, overlying apparent beach deposits, supports this interpretation. In some cases, as in Cycle 5, the beach deposits are not developed or preserved, but this could be due either to erosion and reworking of the beach deposits as the surf-zone migrated inland (e.g. Masters, 1967), or to lack of development of a beach at a distributary mouth. In Cycle 3 the beach deposits are overlain by thinly-interbedded sandstone and siltstone

intercalated with bioturbated sandstone, indicating a return to tidal-flat sedimentation (Lithofacies B). In this case, the surf-zone deposits probably lie further to the seaward (east) of the exposure.

Lithofacies E generally passes gradationally upwards into thin-bedded sandstone and siltstone of Lithofacies B, or may be directly overlain by a channel unit of Lithofacies A type. The contact area is usually marked by an increase in bioturbation, and by the appearance of siltstone layers and channel units. In some of the thin-bedded zones below the measured section, however, another lithofacies resembling the basal Transition Beds appears to be present above Lithofacies E. This probably represents deposition in areas further off-shore than the surf-zone, as suggested for the Transition Beds.

7. Origin of the Cycles

The above interpretations of the various lithofacies suggest that the cycles can be considered in terms of three main environments, viz. a basal distributary channel-tidal flat complex, followed by a lagoon-barrier beach sequence, followed by more marine surf-zone or distributary mouth-bar deposits. The latter in turn give way to tidal flats and channels of the next cycle (Fig. 15). This could be interpreted as a transgression followed by a regression to tidal flats to begin the new cycle. Such a sequence might be due to regular sea-level fluctuations in the manner described by Fischer (1961) and Masters (1967), but another explanation, viz. switching of depositional sites on a delta complex, is also possible.

Somewhat similar cycles, involving alternations of "non-marine" (sandstone-coal) and "marine" (shale-limestone) sequences, are known throughout the world, particularly from the Carboniferous. A comprehensive review of these cycles, and of the theories of origin, has recently been presented by Duff et al. (1967). These authors emphasize the vertical and lateral variability of the cycles, the contemporaneous development of different lithofacies in different areas, and the general lack of correlation of most units beyond local areas. They conclude that most such cycles are best explained in terms of sedimentational mechanisms rather than by the classical tectonic, climatic or eustatic theories.

Previous explanations of Carboniferous cycles in terms of sedimentational phenomena, particularly "delta-switching", had been made by D. Moore (1950, 1959, 1960), Goodlet (1959), Duff and Walton (1962, and Read (1965). These interpretations have been supported by the sub-surface studies of modern deltas, particularly those of the Mississippi (e.g. Coleman and Gagliano, 1964; Scruton, 1960) and the Rhone (Oomkens, 1967; Lagaay and Kopstein, 1964; van Straaten, 1960; Scruton, 1960; Kruit, 1955). The basis of the theory concerns the shifting of the main site of clastic deposition to a new lower area on the delta complex, and the transgression of marine conditions over the abandoned lobe as it subsides below sea-level. Return of the distributary system over the subsiding area begins a new cycle. In the case of the Mississippi the birds-foot pattern of distributaries probably also produces minor cycles due to shifting of distributary arms over inter-distributary bay areas (Coleman and Gagliano, 1964).

The kind of sequence formed will vary greatly according to the kind of delta, the rate of subsidence, the prevailing marine conditions etc., but the general mechanism seems capable of explaining most marine-non-marine cycles without recourse to tectonics or to abrupt climatic or sea-level changes to account for the induction of each cycle.

In the present case it is apparent that the Mississippi model differs in several important aspects from the postulated conditions. Firstly, the Mississippi sediment load consists predominantly of silt and clay, whereas the present deposits are predominantly sandy. Secondly, the Mississippi distributaries are mostly straight (with very pronounced levees) rather than meandering, and hence the channel deposits tend to be laterally restricted (the "bar-finger sands" of Fisk, 1961) rather than typical meander-belt deposits. Thirdly, the tidal influence on the Mississippi delta is relatively slight, and extensive tidal-flat deposits are not developed.

What seems to be required is a sandy delta with meandering distributaries and considerable tidal influence. The Rhone delta fills these requirements to a large extent, and the sub-surface sequences described by Oomkens (1967) have many features in common with the Great Dome Sandstone. The landward part of the Rhone sequence consists

mainly of distributary channel sands lying within coastal plain (brackish to fresh) silts and clays, while towards the seaward edge these deposits are intercalated with barrier sands and marine clays. The "transgressive sequences" of Oomkens, in which lagoonal sediments overlies coastal plain swamps and are in turn overlain by barrier sands followed by marine deposits, appear to be closely analogous to the Great Dome cycles. Several major Quaternary sea-level fluctuations are superimposed on the Rhone sequence, however, and there is some difficulty in separating the effects of these from those of delta-switching.

Analogy can also be made with the modern Niger delta (Allen, 1965c), which is a sandy delta supplied by numerous meandering distributaries connected by tidal areas and mangrove swamps. River-mouth bars separated by beaches form the seaward edge. The sediment discharge is not uniform over the delta but is concentrated in one or two of the major distributaries, the other areas being subject mainly to tidal influences. Although the sub-surface sequences are poorly known, it seems possible that the shifting of the main areas of sediment discharge may result in alternations of fluvial and tidal-marine deposits.

D. UPPER PART OF THE FORMATION

The upper 800 feet or so of the Great Dome Sandstone in the type area consists predominantly of grey, cross-bedded, very quartzose, coarse sandstone and fine conglomerate (Plate 38A). Thin zones of thinly-bedded sandstone and siltstone, usually showing abundant worm burrows, are interbedded with the coarse-grained rocks but generally constitute less than 20% of the section. These thin-bedded rocks are similar to those previously described under Lithofacies B, and are interpreted as being mainly channel-side and tidal flat deposits. Some sections towards the top of the formation consist wholly of conglomeratic sandstone and fine conglomerate and do not show any fining-upwards sequences or thin-bedded rocks. They are generally pink in colour and were possibly deposited mainly by subaerial braided-streams rather than meander-channels. Most of the upper part of the formation is interpreted as a floodplain deposit formed on the upper reaches of a deltaic plain with slight tidal influence.



A. Typical cross-bedded sandstones and fine conglomerates of upper part of Great Dome Sandstone near Reeds Peak.



B. North end of Denison Range showing transition from the Cyclic Facies below to red sandstones of upper part of Great Dome Sandstone (centre) to the thick-bedded conglomerates of Reeds Conglomerate on crest. View looks north.

Towards the northern end of the range (north of Bonds Craig) the upper half of the formation is mostly red in colour and consists almost wholly of cross-bedded coarse sandstones and fine conglomerates (Plate 38B), suggesting that this area was a subaerial floodplain for most of the time. South of this there are interfingering red and grey zones, while at Reeds Peak the main change to red colour coincides with the base of the Reeds Conglomerate.

The contact with the Reeds Conglomerate is transitional through several hundred feet of fine conglomerate and conglomeratic sandstone in most areas, and is usually marked by a gradual change from grey to red colour accompanied by an increase in bedding thickness, an increase in the proportion of conglomerate, and a coarsening of grain-size from pebble to cobble grade. A similar transition occurs in the Bonds Craig area (Plate 38B) except for the colour change. The red colour of the Reeds Conglomerate, the absence of fauna and of fining-upwards sequences, and the presence of scours and channel structures, suggest deposition by braided streams in a non-marine environment, most probably on large alluvial fans.

CHAPTER 4

SUMMARY OF DEPOSITIONAL HISTORY OF DENISON RANGE BASIN

The Denison Range Basin is a small structure forming part of the northern end of the eugeosynclinal Adamsfield Trough. Its development appears to have been controlled largely by the fault system to the west and south which forms the main contact with the Tyennan Geanticline. The nature of the eastern margin of the basin is not known. Most of the filling of the basin occurred in the Upper Cambrian, after the main igneous activity within the trough had ceased. The sediments show a transition from flysch-type turbidites and siltstones at the base to molasse-type sandstones and conglomerates at the top. The marginal position of the basin and the nature of the filling are reminiscent in some respects of the "continental borderland" basins of coastal California (e.g. Shepard and Emery, 1941; Emery; 1960a, b).

Sedimentation within the basin commenced in possibly the late Middle Cambrian with the conglomeratic Trial Ridge Beds. There has unfortunately not been time to study these rocks in detail, and their interpretation in terms of depositional environments is very tentative at this stage. The lower half of the sequence consists mainly of siliceous conglomerates with some intercalated cross-bedded sandstones, and is mostly red in colour in the lower part with alternating zones of red and grey colour towards the top. The lower beds appear to be mainly alluvial fan deposits formed during the earliest stages of basin formation, while the upper beds may be partly shallow marine.

The main upper part of the Trial Ridge Beds consists of turbidites and siltstones intercalated with coarse siliceous conglomerates which appear to have been formed by mass movement of some kind. This part of the sequence represents deeper water marine conditions, and the conglomerates suggest proximity to source and to an active slope area. The sequence is interpreted as an unstable submarine fan deposit formed at the foot of a fault-controlled slope (Fig. 16). The only current directions available from the Trial Ridge Beds are five flute

EVOLUTION OF THE DENISON RANGE BASIN

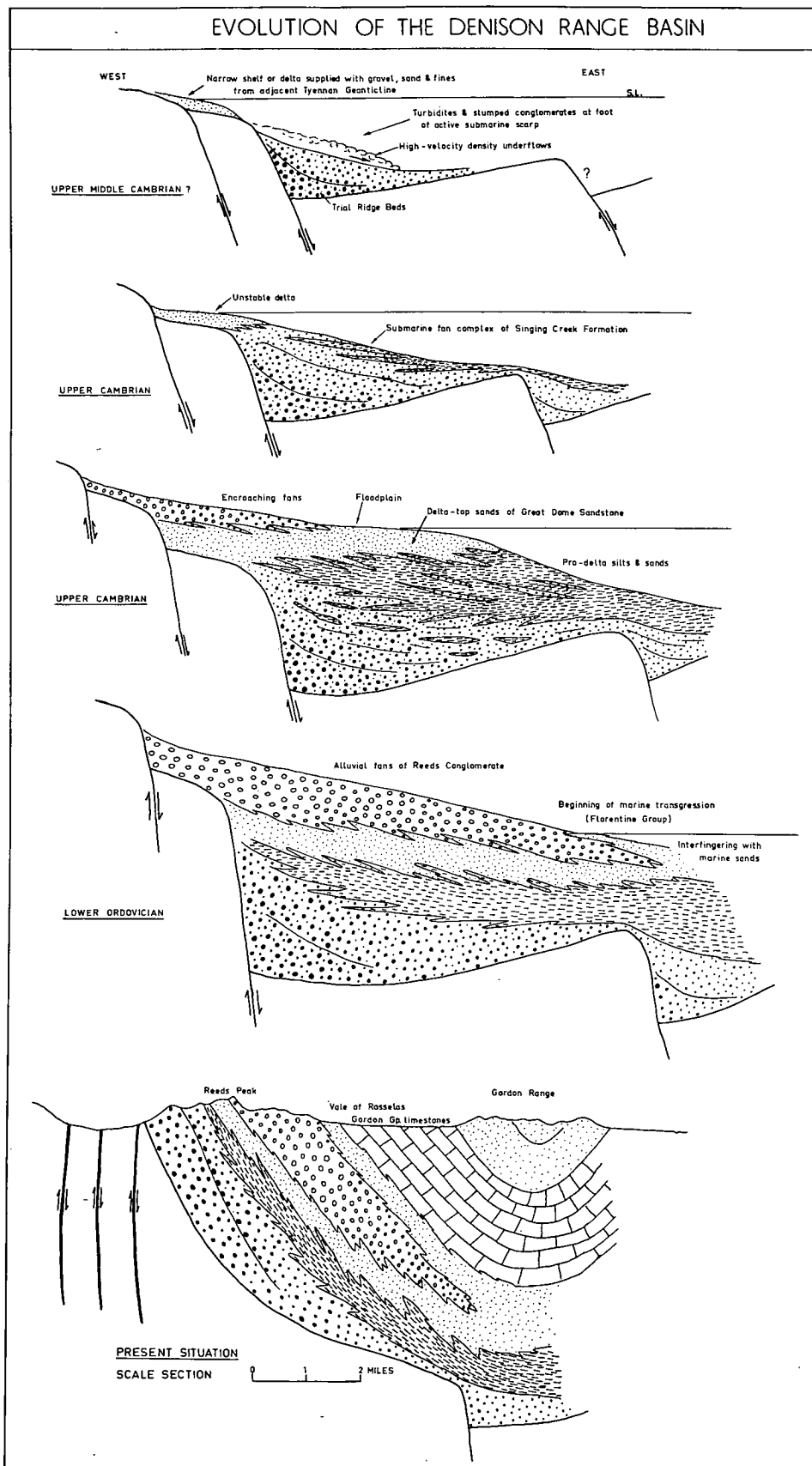


FIGURE 16

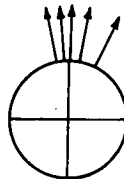
mark orientations (Fig. 17) measured near the northern end of the ridge. These indicate north-flowing currents when the beds are rotated about the present strike. This direction differs somewhat from that of the later formations, and is unusual in that there is no apparent source area for the Precambrian detritus to the south (see Fig. 2). Slight folding and uplift of the Trial Ridge Beds occurred prior to deposition of the Denison Group, resulting in a low-angle unconformity between the two. This movement was possibly related to the intrusion of the Boyes Valley ultramafic body, since this body appears to intrude the Trial Ridge Beds whereas the Denison Group appears to be unconformable above it.

Denison Group sedimentation commenced with a thin breccia-conglomerate derived mainly from the Trial Ridge Beds, followed by 100 feet or so of cross-bedded sandstones and fine conglomerates. These rocks were probably deposited in shallow marine and non-marine environments during the transgression which followed uplift of the Trial Ridge Beds, and pass upwards into marine siltstones, sandstones and conglomerates of the Singing Creek Formation. As previously discussed, this formation appears to be a submarine fan complex formed at the foot of a steep submarine slope. The position of this slope was probably controlled by the same fault system as controlled deposition of the Trial Ridge Beds. Sediment supply to the top of this slope included boulders, cobbles and pebbles of quartzite as well as abundant sand and finer material, suggesting a narrow shelf or possibly a delta area supplied from the adjacent geanticline (Fig. 16).

Deposition on the submarine fans was accomplished mainly by high-velocity density underflows generated on the slope, possibly from slides at the heads of submarine canyons. The denser gravel- and sand-bearing parts of the flows were confined mainly to channels (fan-valleys) and depressions on the fans, where they deposited thick, variable sand and gravel beds. The more diffuse higher parts of the flows overspilled the channels and deposited blanket silts and fine sands over the interfluvial areas. Palaeocurrent directions (Fig. 17) indicate that the surface of the fan sloped towards the ENE in the Flagstone Knoll area. Contemporaneous movements on east-west cross-faults caused considerable slumping of the unconsolidated fan sediments down local slopes oriented

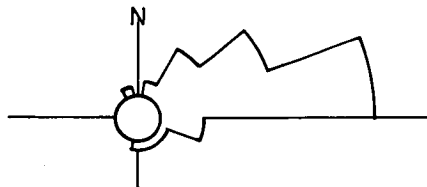
PALAEOCURRENTS - DENISON RANGE BASIN

TRIAL RIDGE BEDS



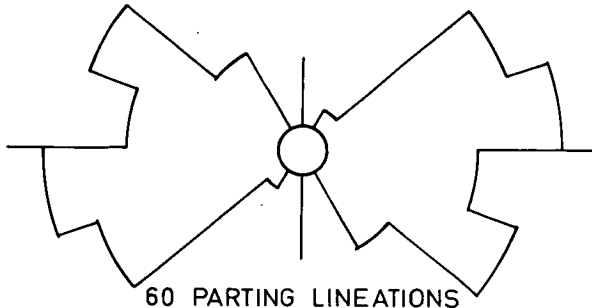
5 FLUTE MARKS

SINGING CREEK FORMATION



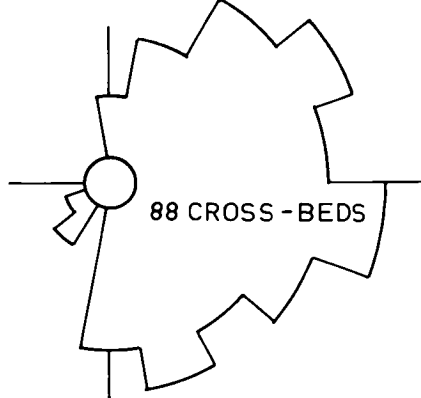
67 CURRENT DIRECTIONS

CYCLIC FACIES - GREAT DOME SANDSTONE



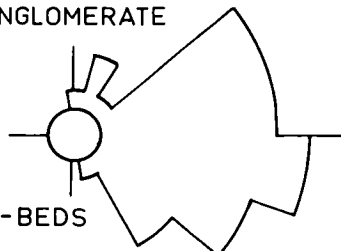
60 PARTING LINEATIONS

UPPER PART GREAT DOME SANDSTONE



88 CROSS - BEDS

REEDS CONGLOMERATE



63 CROSS-BEDS

FIGURE 17

at high angles to the regional slope. The position of these scarps may have controlled to some extent the areas where channels or fan-valleys were developed.

As the basin filled, shallow marine and deltaic facies migrated over the fan deposits. The siltstones which form the upper part of the Singing Creek Formation possibly represent pro-delta deposits of the advancing delta. Most of the Great Dome Sandstone appears to have been deposited on the fluvio-marine floodplain of this delta (Fig. 16). This kind of flysch-paralic transition is very similar to that described from the Upper Carboniferous of northern England by Walker (1966a), where the Mam Tor Sandstones (basin-floor turbidites) pass upwards through the Shale Grit and Grindslow Shales (submarine fan) to the Kinderscout Grit (deltaic-shallow marine).

Lateral shifting of the meandering delta-distributary systems allowed lagoonal, barrier island and shallow marine facies to encroach over parts of the delta at intervals, resulting in alternations of fluvatile-tidal and marine deposits in a cyclic fashion in the lower part of the formation. The sandy nature of the sediments and the absence of vegetation must have promoted meandering, lateral erosion and reworking by the distributaries and tidal channels. One would expect that wind action would also have been pronounced because of the lack of stabilizing vegetation, but evidence for this is difficult to find. Dune ridges and other wind-deposited sediments are apparently absent. Parting lineations measured in the Cyclic Facies, mostly from the surf-zone deposits, show a strong east-west alignment (Fig. 17). This structure is probably wave-produced for the most part, and the east-west alignment suggests a shoreline oriented north-south in agreement with the other palaeocurrents.

Further filling of the basin saw the advance of the upper floodplain over the fluvio-marine lower part, with most of the deposition being accomplished by the distributary channels and only minor accumulations in the inter-distributary tidal-flat areas. Further north, the floodplain was almost entirely non-marine, and most of the sands and gravels were oxidized to red colour. Part of the deposition during this period was by braided streams as well as meander channels. Cross-

bedding directions measured in the upper part of the Great Dome Sandstone (Fig. 17) show a considerable spread, but indicate predominantly easterly-flowing currents. Much of the variation probably reflects the meandering nature of the channels.

To the west, towards the rising Tyennan Geanticline, the upper part of the floodplain probably interfingered with the braided system on the lower reaches of large alluvial fans (Fig. 16). These fans migrated eastward as uplift on the marginal faults continued, and became established over the area in the late Upper Cambrian and early Ordovician, forming the Reeds Conglomerate. The coarseness of these deposits (up to boulder grade), the lack of silt or finer material, the high degree of rounding, the presence of channel structures and abundant cross-bedding, and the absence of any mudflow deposits, indicate deposition largely by powerful braided streams on very large fans. Cross-bedding directions (Fig. 17) again indicate easterly-directed currents on the Denison Range. The thickness of these deposits (maximum of 5000 feet in middle of basin) suggests considerable uplift of the Precambrian source rocks.

To the east, the sandy lower parts of the fans appear to have interfingered with shallow marine sands (e.g. Tim Shea Sandstone), and some intercalations of marine conditions are indicated by horizons of grey, worm-burrowed, flat-bedded sandstone within the conglomerate in a few places. Uplift of the Tyennan Geanticline apparently slowed down in the early Lower Ordovician, and coastal plain alluvial sands were deposited over the conglomerate. The existence of the Denison Range Basin as a separate unit beyond this period is doubtful, and the transgressive sequence which followed appears to largely blanket these early marginal basins. Shallow marine sands and silts of the Florentine Group transgressed over the coastal plain as conditions became more stable, and carbonates of the Gordon Group were deposited as the supply of terrigenous material declined.

BIBLIOGRAPHY

- Allen, J.R.L., 1963. The classification of cross-stratified units, with notes on their origin. Sedimentology, 2, 93-114.
- _____, 1964a. Studies in fluvial sedimentation: six cycles from the Lower Old Red Sandstone, Anglo-Welsh Basin. Sedimentology, 3, 163-198.
- _____, 1964b. Primary current lineation in the Lower Old Red Sandstone (Devonian), Anglo-Welsh Basin. Ibid, 3, 89-108.
- _____, 1965a. Fining-upwards cycles in alluvial successions. L'pool. Manch'r. Geol. J., 4, 229-246.
- _____, 1965b. A review of the origin and characteristics of recent alluvial sediments. Sedimentology, 5, 91-191.
- _____, 1965c. Late Quaternary Niger delta and adjacent areas: sedimentary environments and lithofacies. Am. Assoc. Petroleum Geol., Bull., 49, 547-600.
- _____, 1965d. Upper Old Red Sandstone palaeogeography (Farlovian) in south Wales and the Welsh Borderland. J. Sediment. Petrol., 35, 167-195.
- _____, 1968. On criteria for the continuance of flute marks, and their implications. Geol. Mijnb., 47, 3-16.
- _____, 1969. Erosional current marks on weakly cohesive mud beds. Jour. Sediment. Petrol., 39, 607-623.
- Bagnold, R.A., 1954. Experiments on a gravity-free dispersion of large solid spheres in a Newtonian fluid under shear. Roy. Soc. Lond. Proc., Ser. A, 225, 49-63.
- _____, 1956. The flow of cohesionless grains in fluids. Roy. Soc. Lond., Phil. Trans., Ser. A, 249, 235-297.
- _____, 1962. Auto-suspension of transported sediment: turbidity currents. Roy. Soc. Lond. Proc., Ser. A, 265, 325-319.
- Banks, M.R., 1956. The Middle and Upper Cambrian Series (Dundas Group and its correlates) in Tasmania; pp. 165-212 in El Sistema Cambrico, 2, 20th Int. Geol. Congr., Mexico.
- _____, 1957. The stratigraphy of Tasmanian limestones. Miner. Resour. Tasm., 10, 39-85.

- Banks, M.R., 1961. Age and relationships of Tasmanian fossil faunas and floras. Proc. 9th Pacific Sci. Congr., 1957, 12, 328-339.
- _____, 1962a. Cambrian System; in Geology of Tasmania. Jour. Geol. Soc. Aust., 9, 2, 127-145.
- _____, 1962b. Ordovician System. Ibid, 147-176.
- _____, and Johnson, J.H., 1957. The occurrence of Maclurites and Girvanella in the Gordon Limestone (Ordovician) of Tasmania. Jour. Paleont., 31, 632-640.
- Bassett, D.A. and Walton, E.K., 1959. The Hell's Mouth Grits; Cambrian greywackes in St. Tudwal's Peninsula, North Wales. Quart. J. Geol. Soc. Lond., 116, 85-110.
- Bernard, H.A. and Major, C.F., 1963. Recent meander belt deposits of the Brazos River: an alluvial "sand" model (Abstract). Am. Assoc. Petrol. Geol. Bull., 47, 350.
- Blatt, H. and Christie, J.M., 1963. Undulatory extinction in quartz of igneous and metamorphic rocks and its significance in provenance studies of sedimentary rocks. Jour. Sediment. Petrol., 33, 559-579.
- Blissenbach, E., 1954. Geology of alluvial fans in semi-arid regions. Geol. Soc. Amer., Bull., 65, 175-190.
- Bluck, B.J., 1965. The sedimentary history of some Triassic conglomerates in the Vale of Glamorgan, South Wales. Sedimentology, 4, 225-245.
- Bouma, A.H., 1962. Sedimentology of some Flysch Deposits. A Graphic Approach to Facies Interpretation. Elsevier, Amsterdam, 168pp.
- _____ and Shepard, F.P., 1964. Large rectangular cores from submarine canyons and fan valleys. Am. Assoc. Petroleum Geol., 48, 225-231.
- Briggs, L.I. and Middleton, G.V., 1965. Hydromechanical principles of sediment structure formation. Soc. Econ. Pal. Min. Spec. Pub. No.12, 5-16.
- Brown, I.A., 1948. Lower Ordovician brachiopods from Junee district, Tasmania. Jour. Paleont., 22, 35-39.

- Buffington, E.C., Moore, D.G., Dill, R.F. and Vernon, J.W., 1967. From shore to abyss: near-shore transport, slope deposition and erosion, canyon transport, and deep-basin sedimentation. (Abstract). Am. Assoc. Petrol. Geol. Bull., 51, 456.
- Bull., W.B., 1963. Alluvial fan deposits in western Fresno County, California. J. Geol., 71, 243-251.
- Burns, K.L., 1963. The tectonic history of the Dial Range area, Tasmania. Unpub. Ph.D. Thesis, Geol. Dep't., Univ. of Tas.
- Campana, B., Dickinson, S.B., King, D. and Matheson, R.S., 1958. The mineralized rift valleys of Tasmania. In Stillwell Anniversary Volume, Aust. Inst. Min. Engrs., 41-60.
- _____, Dickinson, S.B. and King, D., 1959. The mineralized rift valleys of Tasmania. Proc. Aust. Inst. Min. Engrs., 191, 198-202.
- _____ and King, D., 1963. Palaeozoic tectonism, sedimentation and mineralization in West Tasmania. Jour. Geol. Soc. Aust., 10, 1, 1-53.
- Carey, S.W., 1947. Report of the Government Geologist. Rept. Dir. of Mines Tasm. for 1945, 21-29.
- _____, 1953. Geological structure of Tasmania in relation to mineralization. 5th Emp. Min. Metall. Congr., 1, 1108-1128.
- _____, 1960. Tasmanian University Seismic Net. Publ. Geol. Dep. Univ. Tasm., 84.
- _____ and Banks, M.R., 1954. Lower Palaeozoic unconformities in Tasmania. Pap. Roy. Soc. Tasm., 88, 245-269.
- Chapman, F., 1919. On the occurrence of Tetradium in Gordon River Limestone, Tasmania. Rec. Geol. Surv. Tasm., 5.
- Chenowith, P.A. 1952. Statistical methods applied to Trentonian stratigraphy in New York. Bull. Geol. Soc. Amer., 63, 521-560.
- Coleman, J.M. and Gagliano, S.M., 1964. Cyclic sedimentation in the Mississippi River deltaic plain. Gulf Coast Assoc'd. Geol. Soc's, Trans., 14, 67-80.

- Coleman, J.M. and Gagliano, S.M., 1965. Sedimentary structures: Mississippi River delatic plain. Soc. Econ. Pal. Min. Spec. Pub. No.12, 133-148.
- Conybeare, C.E.B. and Crook, K.A.W., 1968. Manual of Sedimentary Structures. Bur. Min. Resour. Geol. Geoph., Austr., Bull. No.102, 327pp.
- Corbett, K.D., 1964. Geology of the Florentine Valley area. Unpublished Honours thesis, Geology Dep't., University of Tasmania.
- Craig, G.V. and Walton, E.K., 1962. Sedimentary structures and palaeo-current directions from the Silurian rocks of Kirkcudbrightshire. Trans. Edinb. Geol. Soc., 19, 100-119.
- Crook, K.A.W., 1960. Classification of arenites. Amer. J. Sci., 258, 419-428.
- _____, 1961. Stratigraphy of the Parry Group (Upper Devonian-Lower Carboniferous), Tamworth-Nundle district, N.S.W. J. Roy. Soc. N.S.W., 94, 189-207.
- Crowell, J.C., 1955. Directional current structures from the Pre-alpine flysch, Switzerland. Geol. Soc. Amer. Bull., 66, 1351-1384.
- _____, 1957. Origin of pebbly mudstones. Ibid, 68, 993-1009.
- _____, Hope, R.A., Kahle, J.E., Ovenshine, A.T. and Sans, R.A., 1966. Deep-water sedimentary structures of the Pliocene Pico Formation, Santa Paula Creek, Ventura Basin, California. Calif. Dep't. Nat. Resour., Div. Mines, Spec. Rep't., 89, 40pp.
- Cummins, W.A., 1962. The greywacke problem. L'pool Manch'r Geol. Jour., 3, 51-72.
- Daly, R.A., 1936. Origin of submarine canyons. Am. J. Sci., 31, 401-420.
- Davies, J.L., 1965. Landforms. In Atlas of Tasmania (Ed. J.L. Davies). Lands and Surveys Department, Hobart.
- Denny, C.S., 1965. Alluvial fans in the Death Valley region, California, Nevada. U.S. Geol. Surv. Prof. Pap. 466, 62pp.
- Dill, R.F., Dietz, R.S. and Stewart, H.B., 1954. Deep-sea channels and delta of the Monterey submarine canyon. Bull. Geol. Soc. Amer., 65, 191-193.

- Dott, R.H. Jr., 1963. Dynamics of subaqueous gravity depositional processes. Am. Assoc. Petroleum Geol., Bull., 47, 104-128.
- _____, 1964. Wacke, graywacke and matrix - what approach to immature sandstone classification? Jour. Sedimentary Petrology, 34, 625-632.
- Duff, P.McL.D. and Walton, E.K., 1962. Statistical basis for cyclothems: a quantitative study in the East Pennine Coalfield. Sedimentology, 1, 235-255.
- _____, Hallam, A. and Walton, E.K., 1967. Cyclic Sedimentation. Elsevier, Amsterdam. 280pp.
- Dzulynski, S., Ksiazkiewicz, M. and Kuenen, P.H. 1959. Turbidites in flysch of the Polish Carpathians. Geol. Soc. Amer. Bull., 70, 1089-1118.
- _____ and Sanders, J.E., 1962. Current marks on firm mud bottoms. Trans. Conn. Acad. Arts Sci., 42, 57-96.
- _____ and Walton, E.K., 1965. Sedimentary Features of Flysch and Greywackes. Elsevier, Amsterdam.
- Elliston, J., 1953. Platinoids in Tasmania. 5th Emp. Min. Metall. Congr., 1, 1250-1254.
- Emery, K.O., 1945. Entrapment of air in beach sand. J. Sediment. Petrol., 15, 39-49.
- _____, 1950. Contorted Pleistocene strata at Newport Beach, California. J. Sediment. Petrol., 20, 111-115.
- _____, 1960a. The Sea off Southern California. A Modern Habitat of Petroleum. John Wiley and Sons, New York. 366pp.
- _____, 1960b. Basin plains and aprons off southern California. J. Geol., 68, 464-479.
- _____ and Stevenson, R.E., 1950. Laminated beach sand. J. Sediment. Petrol., 20, 220-223.
- Emrich, G.H. and Wobber, F.J., 1963. A rapid visual method for estimating sedimentary parameters. Jour. Sediment. Petrol., 33, 831-843.
- Etheridge, R., 1904. Trilobite remains collected in the Florentine Valley, western Tasmania, by Mr. T. Stephens, M.A. Rec. Aust. Mus., 5, 89-101.

- Evans, G., 1965. Intertidal flat sediments and their environments of deposition in the Wash. Quart. J. Geol. Soc. Lond., 121, 209-245.
- Fischer, A.G., 1961. Stratigraphic record of transgressing seas in light of sedimentation on Atlantic coast of New Jersey. Am. Assoc. Petrol. Geol. Bull., 45, 1656-1666.
- Fisk, H.N., 1961. Bar-finger sands of Mississippi Delta. In Geometry of Sandstone Bodies, A.A.P.G., Tulsa, 29-52.
- Folk, R.L., 1951. Stages in textural maturity in sedimentary rocks. Jour. Sediment. Petrol., 21, 127-130.
- _____, 1954. The distinction between grain size and mineral composition in sedimentary rock nomenclature. J. Geol., 62, 344-359.
- _____, 1961. Petrology of Sedimentary Rocks. Hemphills Book Store, Austin, Texas.
- Frazier, D.E. and Osanik, A., 1961. Point-bar deposits, Old River locksite, Louisiana. Gulf Coast Assoc'd. Geol. Soc's. Trans., 11, 212-237.
- Gebert, H., 1965. Geology and mineralization of the Adamsfield area. Broken Hill Prop. Coy. Ltd., unpub. rep't.
- _____, 1966. Notes on geological reconnaissance mapping on Map Sheet Wedge E.L. 13/65, S.W. Tasmania. Ibid, unpub. rep't.
- Gee, R.D., 1967. The tectonic evolution of the Rocky Cape Geanticline in northwest Tasmania. Unpub. Ph.D. thesis, Geol. Dep't., Univ. of Tasm.
- Gilbert, C.M., 1954. Sedimentary rocks, in Williams, Turner and Gilbert, Petrography. W.H. Freeman, New York, 406pp.
- Goodlet, G.A., 1959. Mid-Carboniferous sedimentation in the Midland Valley of Scotland. Trans. Edinb. Geol. Soc., 17, 217-240.
- Gorsline, D.S. and Emery, K.O., 1959. Turbidity current deposits in San Pedro and Santa Monica basins off southern California. Bull. Geol. Soc. Amer., 70, 279-290.

- Gould, C., 1866. On the position of the Gordon Limestone relatively to other Palaeozoic formations. Pap. Roy. Soc. Tasm., 4, 27-29.
- Gregory, H.E., 1915. The formation and distribution of fluviatile and marine gravels. Amer. J. Sci., 39, 487-508.
- Gregory, M.R., 1969. Sedimentary features and penecontemporaneous slumping in the Waitemata Group, Whangaparaoa Peninsula, North Auckland, New Zealand. N.Z. J. Geol. Geoph., 12, 248-282.
- Groves, D.I., 1968. The cassiterite-sulphide deposits of western Tasmania. Unpub. Ph.D. thesis, Geol. Dep't., Univ. of Tasm.
- Hall, W.D.M., 1967. Report on geological investigations, S.W. Tasmania, E.L. 13/65. Broken Hill Prop. Coy. Ltd., unpub. rep't.
- Hand, B.M. and Emery, K.O., 1964. Turbidites and topography of the north end of the San Diego trench, California. J. Geol., 72, 526-542.
- Hantzschel, W., 1962. Trace fossils and problematica. In Treatise on Invertebrate Paleontology, Part W. Miscellaneous. (Editor R.C. Moore), 177-245.
- Happ, S.C., Rittenhouse, G. and Dobson, G.C., 1940. Some aspects of accelerated stream and valley sedimentation. U.S. Dep't. Agric., Tech. Bull. 695, 134pp.
- Harms, J.C., Mackenzie, D.B. and McCubbin, D.G., 1963. Stratification in modern sands of the Red River, Louisiana. J. Geol., 71, 566-580.
- _____ and Fahnestock, R.K., 1965. Stratification, bed forms and flow phenomena, with an example from the Rio Grande. Soc. Econom. Pal. Min. Spec. Publ. No.12, 84-115.
- Harrison, S.C. and Jacka, A.D., 1967. Depositional environment of Cherry Canyon sandstone tongue, Last Chance Canyon, New Mexico (Abstract). Am. Assoc. Petrol. Geol. Bull., 51, 468.

- Hayes, M.O., Anan, F.S. and Bozeman, R.N., 1969. Sediment dispersal trends in the littoral zone; a problem in palaeogeographic reconstruction. In Coastal Environments of Northeastern Massachusetts and New Hampshire. Coastal Research Group, Univ. of Mass., Field Trip Guidebook, 290-315.
- _____, Boothroyd, J.C. and Hine, A.C., 1969. Diagnostic primary structure of estuarine sand bodies (Abstract). Ibid, 415.
- Henderson, Q.J., 1939. A brief note on the geology of the country in the vicinity of the old Humboldt Mine. Tasm. Dep't. Mines Rep't., unpub.
- Hills, C.L., 1921. Geological conditions affecting water conservation in the Florentine and Gordon Valleys. Tasm. Dep't. Mines Rep't., unpub.
- _____ and Carey, S.W. 1949. Geology and Mineral Industry; in Handbook for Tasmania. Aust. Ass. Advanc. Sci., Hobart.
- Hoyt, J.H., 1962. High-angle beach stratification, Sapelo Island, Georgia. J. Sediment. Petrol., 32, 309-311.
- Hsu, K.J., 1959. Flute and groove casts in the Pre-alpine flysch, Switzerland. Amer. J. Sci., 257, 529-536.
- _____, 1964. Cross-lamination in graded bed sequences. Jour. Sediment. Petrol., 34, 379-388.
- Hubert, J.F., 1964. Textural evidence for deposition of many North Atlantic deep-sea sands and silts by ocean-bottom currents rather than turbidity currents. J. Geol., 72, 757-785.
- _____, 1966. Sedimentary history of Upper Ordovician geosynclinal rocks, Girvan, Scotland. Jour. Sediment. Petrol., 36, 677-699.
- _____, Scott, K.M. and Walton, E.K., 1966. Composite nature of Silurian flysch sandstones shown by groove moulds on intra-bed surfaces, Peeblesshire, Scotland. Jour. Sediment. Petrol., 36, 237-241.

- Hughes, T.D., 1952. Copper prospects, Mt. Mueller mine. Tasm. Dep't. Mines Rep't., unpub.
- Imbrie, J. and Buchanan, H., 1965. Sedimentary structures in modern carbonate sands of the Bahamas. Soc. Econ. Pal. Min. Spec. Publ. No.12, 149-172.
- Jacka, A.D. and St.Germain, L.C., 1967. Deep-sea fans in Permian Delaware Mountain Group, Delaware Basin, West Texas and New Mexico (Abstract). Am. Assoc. Petrol. Geol. Bull., 51, 471.
- Jago, J.B., 1966. The geology of the Maydena Range. Unpublished Honours thesis, Geology Dep't., Univ. of Tasm.
- Jahns, R.H., 1947. Geological features of the Connecticut Valley, Mass., as related to recent floods. U.S. Geol. Surv. Water-Supp. Pap. 996, 158pp.
- Jennings, I.B., 1955. Geology of portion of the Middle Derwent area. Pap. Roy. Soc. Tasm., 89, 169-190.
- _____, Noldart, A.J. and Williams, E., 1967. Geology and mineral resources of Tasmania. Bull. Geol. Surv. Tasm., No.50.
- Jennings, J.N. and Banks, M.R., 1958. The Pleistocene glacial history of Tasmania. J. Glaciol., 3, 298-303.
- Johnson, K.G. and Friedman, G.M., 1969. The Tully clastic correlatives (Upper Devonian) of New York State: a model for recognition of alluvial, dune (?), tidal, near-shore (bar and lagoon), and offshore sedimentary environments in a tectonic delta complex. J. Sediment. Petrol., 39, 451-485.
- Johnston, R.M., 1888. Systematic Account of the Geology of Tasmania. Gov't. Printer, Hobart.
- Jones, O.T., 1939. The geology of the Colwyn Bay district: a study of submarine slumping during the Salopian period. Quart. J. Geol. Soc. Lond., 95, 335-376.
- Jopling, A.V., 1965. Hydraulic factors controlling the shape of laminae in laboratory deltas. Jour. Sediment. Petrol., 35, 777-791.

- Kingma, J.T., 1958. The Tongaporutuan sedimentation in central Hawke's Bay. N.Z. J. Geol. Geoph., 1, 1-30.
- _____, 1960. The tectonic significance of graded bedding in geosynclinal sedimentary systems. 21st Int. Geol. Congr., Copenhagen, Pt.XXI, 205-214.
- Kingsbury, C.J.R., 1961. Report on Mt. Wedge area E.L. 1/59. Lyell-E.Z. Explorations, Ltd., unpub. rep't.
- Klein, G.D., 1963a. Analysis and review of sandstone classifications in the North American literature 1940-1960. Geol. Soc. Amer. Bull., 74, 555-576.
- _____, 1963b. Bay of Fundy intertidal zone sediments. J. Sediment. Petrol., 33, 844-854.
- _____, 1965. Dynamic significance of primary structures in the Middle Jurassic Great Oolite Series, southern England. Soc. Econ. Pal. Min. Spec. Pub. No.12, 173-191.
- _____, 1966. Dispersal and petrology of the Stanley-Jackfork boundary, Ouachita fold belt, Arkansas and Oklahoma. Am. Assoc. Petrol. Geol. Bull., 50, 308-326.
- Kobayashi, T., 1936. Three contributions to Cambro-Ordovician stratigraphy and faunas. Jap. Jour. Geol. Geogr.
- _____, 1940. Lower Ordovician fossils from Junee, Tasmania. Pap. Roy. Soc. Tasm. (1939), 61-66.
- Kruit, C., 1955. Sediments of the Rhone delta. Verkh. Kon. Ned. Geol. Mijnb. Genoot., Geol. Ser., 15, 357-514.
- Krynine, P.D., 1948. Megascopic study and field classification of sedimentary rocks. J. Geol., 56, 130-165.
- _____, 1950. Petrology, stratigraphy and origin of Triassic sedimentary rocks of Connecticut. Conn. State Geol. Nat. Hist. Surv., Bull., 73, 247pp.
- Ksiazkiewicz, M., 1961. Life conditions in flysch basins. Ann. Soc. Geol. Pologne, 31, 3-21.
- Kuenen, P.H., 1948. Slumping in the Carboniferous rocks of Pembrokeshire. Quart. J. Geol. Soc. Lond., 104, 365-380.
- _____, 1951. Properties of turbidity currents of high density. Soc. Econ. Pal. Min. Spec. Pub. No.2, 14-33.

- Kuenen, P.H., 1953. Graded bedding, with observations on Lower Palaeozoic rocks of Britain. Verh. Kon. Ned. Akad. Wetters. Amsterd., Afd. Nat., 20, 1-47.
- _____, 1964. Deep-sea sands and ancient turbidites. In Bouma and Brouwer (Editors), Turbidites, Developments in Sedimentology 3; Elsevier, Amsterdam, 264pp.
- _____, 1966a. Matrix of turbidites: experimental approach. Sedimentology, 7, 267-297.
- _____, 1966b. Experimental turbidite lamination in a circular flume. J. Geol., 74, 523-545.
- _____, 1967. Emplacement of flysch-type sand beds. Sedimentology, 9, 203-244.
- _____ and Migliorini, C.I., 1950. Turbidity currents as a cause of graded bedding. J. Geol., 58, 91-127.
- Lagaay, R. and Kopstein, F.P.H.W., 1964. Typical features of a fluvio-marine offlap sequence. In Deltaic and Shallow Marine Deposits, (Editor Van Straaten), Elsevier, Amsterdam, 216-226.
- Land, L.S. and Hoyt, J.H., 1966. Sedimentation in a meandering estuary. Sedimentology, 6, 191-207.
- _____, Mackenzie, F.T. and Gould, S.J., 1967. Pleistocene history of Bermuda. Bull. Geol. Soc. Amer., 78, 993-1006.
- Lane, D.W., 1963a. Cross-stratification in San Bernard River, Texas, point-bar deposits. J. Sediment. Petrol., 33, 350-354.
- _____, 1963b. Sedimentary environments in Cretaceous Dakota Sandstone in northwestern Colorado. Am. Assoc. Petrol. Geol. Bull., 47, 229-256.
- Leitch, E.C. and Mayer, W., 1969. Mass-movement phenomena in an Upper Carboniferous greywacke-argillite sequence in north-eastern New South Wales. N.Z. J. Geol. Geoph., 12, 156-171.
- Lewis, A.N., 1924. Notes on a geological reconnaissance of Mt. Anne and Weld River Valley, south-west Tasmania. Pap. Roy. Soc. Tasm. (1923), 9-24.
- _____, 1940. Geology of the Tyenna Valley. Pap. Roy. Soc. Tasm. (1939), 33-59.

- Macar, P., 1948. Les pseudonodules du Famennien et leur origine. Ann. Soc. Geol. Belg., 72, 47-74.
- Marschalko, R., 1961. Sedimentologic investigation of marginal lithofacies in flysch of Central Carpathians. Geol. Prace, Bratislava, 60, 197-232.
- _____, 1963. Sedimentary slump folds and the depositional slope (flysch of the central Carpathians). Geol. Prace, Zpravy, 28, 161-165.
- _____, 1964. Sedimentary structures and palaeocurrents in the marginal lithofacies of the central Carpathian flysch. In Turbidites, Editors Bouma and Brouwer; Elsevier, Amsterdam. 264pp.
- Masters, C.D., 1967. Use of sedimentary structures in determination of depositional environments, Mesaverde Formation, Williams Fork Mountains, Colorado. Am. Assoc. Petrol. Geol. Bull., 51, 2033-2043.
- McBride, E.F. and Yeakel, L.S., 1963. Relationship between parting lineation and rock fabric. Jour. Sediment. Petrol., 33, 779-782.
- _____ and Kimberly, J.E., 1963. Sedimentology of Smithwick Shale (Pennsylvanian), eastern Llano region, Texas. Am. Assoc. Petrol. Geol. Bull., 47, 1840-1854.
- McKee, E.D., 1957. Primary structures in some recent sediments. Am. Assoc. Petrol. Geol. Bull., 41, 1704-1747.
- Menard, H.W., 1955. Deep-sea channels, topography and sedimentation. Am. Assoc. Petrol. Geol. Bull., 39, 236-255.
- _____, 1960. Possible pre-Pleistocene deep-sea fans off central California. Bull. Geol. Soc. Amer., 71, 1271-1278.
- _____ and Ludwick, J.C., 1951. Applications of hydraulics to the study of marine turbidity currents. Soc. Econ. Pal. Min. Spec. Pub. No.2, 2-13.
- Middleton, G.V., 1965. Antidune cross-bedding in a large flume. Jour. Sediment. Petrol., 35, 922-927.
- _____, 1966a. Small-scale models of turbidity currents and the criteria for autosuspension. Ibid, 36, 202-208.

- Moore, D., 1958. The Yoredale Series of Upper Wensleydale and adjacent parts of northwest Yorkshire. Proc. Yorksh. Geol. Soc., 31, 91-148.
- _____, 1959. Role of deltas in the formation of some British Lower Carboniferous cyclothems. J. Geol., 67, 522-539.
- _____, 1960. Sedimentation units in sandstones of the Yoredale Series (Lower Carboniferous) of Yorkshire, England. J. Sediment. Petrol., 30, 218-227.
- Moss, A.J., 1962. The physical nature of common sandy and pebbly deposits. Part 1. Amer. J. Sci., 260, 337-373.
- _____, 1963. The physical nature of common sandy and pebbly deposits. Part 2. Ibid, 261, 297-343.
- Murphy, M.A. and Schlanger, S.O., 1962. Sedimentary structures in Ilhas and Sao Sebastiao Formations (Cretaceous), Reconcavo Basin, Brazil. Am. Ass. Petrol. Geol. Bull., 46, 457-477.
- Natland, M.L. and Kuenen, P.H., 1951. Sedimentary history of the Ventura Basin, California, and the action of turbidity currents. Soc. Econ. Pal. Min. Spec. Pub. No.2, 76-107.
- Newell, N.D., Rigby, J.K., Fischer, A.G., Whiteman, A.J., Hickox, J.E. and Bradley, J.S., 1953. The Permian Reef Complex of the Guadalupe Mountains Region, Texas and New Mexico. Freeman and Coy., San Francisco. 236pp.
- Nordin, C.F., 1963. A preliminary study of sediment transport parameters, Rio Puerco near Bernado, New Mexico. U.S. Geol. Surv. Prof. Pap. 462C, 20pp.
- Nye, P.B., 1927. Gold prospect at Adamsfield. Tasm. Dept. Mines Rep't. (unpub.).
- _____, 1929. The osmiridium deposits of the Adamsfield district. Bull. Geol. Surv. Tasm., 39, 70pp.
- Okada, H., 1966. Non-greywacke "turbidite" sandstones in the Welsh geosyncline. Sedimentology, 7, 211-232.
- Oomkens, E., 1967. Depositional sequences and sand distribution in a deltaic complex. Geol. en Mijnb., 46, 265-279.
- _____, and Terwindt, J.H.J., 1960. Inshore estuarine sediments in the Haringvliet (Netherlands). Ibid, 39, 701-710.

- Opik, A.A., 1951. Notes on the stratigraphy and palaeontology of Cambrian, Ordovician and Silurian rocks of Tasmania. Bur. Min. Resour. Aust. Rec., 1951-55.
- Otvos, E.G. Jr., 1964. Observation of beach cusp and beach ridge formation on the Long Island Sound. J. Sediment. Petrol., 34, 554-560.
- Packham, G.H., 1954. Sedimentary structures as an important feature in the classification of sandstones. Amer. J. Sci., 252, 466-476.
- Passega, R., 1953. Sedimentary trends, Colorado member of Oficina Formation, San Roque, Anzoategui, Venezuela. Am. Assoc. Petrol. Geol. Bull., 37, 331-339.
- _____, 1954. Turbidity currents and petroleum exploration. Am. Assoc. Petrol. Geol. Bull., 38, 1871-1887.
- Pepper, J.F., De Witt, W. Jr. and Demarest, D.F., 1954. Geology of the Bedford Shale and Berea Sandstone in the Appalachian Basin. U.S. Geol. Surv. Prof. Pap. 259, 109pp.
- Pettijohn, F.J., 1957. Sedimentary Rocks. Harper, New York. 718pp.
- Pike, G., 1964. Geology of the Valentines Peak area. Unpub. Honours thesis, Geol. Dep't., Univ. of Tasm.
- Potter, P.E., 1955. The petrology and origin of the Lafayette Gravel. J. Geol., 63, 1-38.
- _____ and Pettijohn, F.J., 1963: Paleocurrents and Basin Analysis. Springer, Berlin. 296pp.
- Powell, C.McA., 1967. Studies in folding. Unpub. Ph.D. thesis, Geol. Dep't., Univ. of Tasm.
- _____, 1969. Polyphase folding in Precambrian low-grade metamorphic rocks, Middle Gordon River, southwestern Tasmania. Pap. Roy. Soc. Tasm., 103, 47-52.
- Radomski, A., 1960. Remarks on sedimentation of shales in flysch deposits. Bull. Acad. Polon. Sci., Ser. Sci. Geol. Geogr., 8, 123-129.
- Rattigan, J.H., 1967. Depositional, soft sediment and post-consolidation structures in a Palaeozoic aqueoglacial sequence. Jour. Geol. Soc. Aust., 14, 5-18.

- Read, W.A. 1965. Shoreward facies changes and their relation to cyclical sedimentation in part of the Namurian east of Stirling, Scotland. Scott. J. Geol., 1, 69-92.
- Reid, A.M., 1921. Osmiridium in Tasmania. Bull. Geol. Surv. Tasm., 34.
 _____, 1925. Adams River osmiridium field. Tasm. Dep't. Mines Rep't. (unpub.).
- Reineck, H.E., 1964. Layered sediments of tidal flats, beaches and shelf bottoms. In Lauff, G.H. (Editor), Proceedings Conference on Estuaries, Washington, D.C. Am. Assoc. Adv. Sci., 191-205.
- _____ and Wunderlich, F., 1968. Classification and origin of flaser and lenticular bedding. Sedimentology, 11, 99-104.
- Rigby, J.K., 1958. Mass movements in Permian rocks of Trans-Pecos, Texas. Jour. Sediment. Petrol., 28, 298-315.
- Rizzini, A. and Passega, R., 1964. Evolution de la sedimentation et orogenese, vallee du Santerno Apennin septentrional. In Turbidites, Editors Bouma and Brouwer. Elsevier, Amsterdam. 264pp.
- Rubenach, M.J., 1967. The Serpentine Hill Complex. Unpub. Honours thesis, Geol. Dep't., Univ. of Tasm.
- Sanders, J.E., 1960. Origin of convoluted laminae. Geol. Mag., 97, 409-421.
- _____, 1965. Primary sedimentary structures formed by turbidity currents and related resedimentation mechanisms. Soc. Econ. Pal. Min., Spec. Pub. No.12, 192-219.
- Scott, B., 1959. The mineralized rift valleys of Tasmania: discussion and contributions. Proc. Aust. Inst. Min. Engrs., 191, 191-202.
- _____, 1960. Lower Palaeozoic unconformities in south-western Tasmania. Pap. Roy. Soc. Tasm., 94, 103-110.
- Scott, K.M., 1966. Sedimentology and dispersal pattern of a Cretaceous flysch sequence, Patagonian Andes, southern Chile. Am. Assoc. Petrol. Geol. Bull., 50, 72-107.

- Scruton, P.C., 1960. Delta building and the deltaic sequence. In Recent Sediments, North-West Gulf of Mexico, A.A.P.G., Tulsa, 82-102.
- Selley, R.C., Shearman, D.J., Sutton, J. and Watson, J., 1963. Some underwater disturbances in the Torridonian of Skye and Raasay. Geol. Mag., 100, 224-243.
- Shepard, F.P. and Emery, K.O., 1941. Submarine topography off the California coast: canyons and tectonic interpretation. Geol. Soc. Amer. Spec. Pap. No.31, 171pp.
- _____ and Einsele, G., 1962. Sedimentation in San Diego trough and contributing submarine canyons. Sedimentology, 1, 81-133.
- Simons, D.B., Richardson, E.V. and Nordin, C.F. Jr., 1965. Sedimentary structures generated by flow in alluvial channels. Soc. Econ. Pal. Min., Spec. Publ. No.12, 34-52.
- Smith, E., 1957. Lexicon of the stratigraphy of Tasmania. Bur. Min. Resour. Aust., Canberra.
- Soliman, S.M., 1964. Primary structures in a part of the Nile delta beach sand. In Deltaic and Shallow Marine Deposits (Editor van Straaten). Elsevier, Amsterdam, 379-387.
- Solomon, M., 1962. The tectonic history of Tasmania. In Geology of Tasmania. Jour. Geol. Soc. Aust., 9, 2, 311-339.
- _____, 1964. The spilite-keratophyre association of west Tasmania and the ore deposits at Mt. Lyell, Rosebery and Hercules. Unpub. Ph.D. thesis, Geol. Dep't., Univ. of Tasm.
- Sorauf, J.E., 1965. Flow rolls in Upper Devonian rocks of south-central New York State. J. Sediment. Petrol., 35, 553-563.
- Spencer, D.W., 1963. The interpretation of grain size distribution curves of clastic sediments. Jour. Sediment. Petrol., 33, 180-190.
- Spry, A.H., 1957. The Precambrian dolomites of Tasmania. Miner. Resour. Tasm., 10, 32-38.
- _____, 1962. The Precambrian rocks. In Geology of Tasmania. Jour. Geol. Soc. Aust., 9, 2, 107-126.

- Stanley, D.J., 1963a. Vertical petrographic variability in Annot Sandstone turbidites: some preliminary observations and generalizations. Jour. Sediment. Petrol., 33, 783-788.
- _____, 1963b. Non-turbidites in flysch-type sequences. Geol. Soc. Amer. Spec. Pap. 76, 155-156.
- Stauffer, P.H., 1967. Grain flow deposits and their implications, Santa Ynez Mountains, California. Jour. Sediment. Petrol., 37, 487-508.
- Stoneley, R., 1957. On turbidity currents. Verh. Kon. Ned. Geol. Mijnb. Genoot., Geol. Ser., 18, 279-285.
- Sullwold, H.H., 1960. Tarzana fan, deep submarine fan of late Miocene age, Los Angeles County, California. Am. Assoc. Petrol. Geol. Bull., 44, 433-457.
- _____, 1961. Turbidites in oil exploration. In Geometry of Sandstone Bodies. A.A.P.G., Tulsa, Oklahoma, 63-81.
- Sundborg, A., 1956. The River Klaralven. A study of fluvial processes. Geografiska Annal., 38, 127-316.
- Swann, D.H., Fisher, R.W. and Walters, M.J., 1959. Visual estimates of grain size distribution in some Chester sandstones. Illin. Geol. Surv. Circ. 280, 43pp.
- Taylor, B.L., 1955. Asbestos in Tasmania. Miner. Resour. Tasm., 9.
- Teichert, C., 1947. Early Ordovician cephalopods from Adamsfield, Tasmania. Jour. Paleont., 21, 420-428.
- _____ and Glenister, B.F., 1953. Ordovician and Silurian cephalopods from Tasmania, Australia. Bull. Amer. Paleont., 34, 144.
- Terzhagi, K., 1956. Varieties of submarine slope failures. Proc. Texas Conf. Soil Mech. Found. Engr., 8th, 52, 41pp.
- Thomas, D.E., 1945. Report of the Government Geologist. Rep't. Dir. of Mines, Tasm. for 1943, 21-23.
- _____, 1948. A critical review of the Lower Palaeozoic succession in Tasmania. Proc. Roy. Soc. Vict., 59, 23-52.
- _____, 1960. The zonal distribution of Australian graptolites. Jour. Roy. Soc. N.S.W., 94, 1-58.

- Thompson, W.O., 1937. Original structures of beaches, bars and dunes. Bull. Geol. Soc. Amer., 48, 723-752.
- Trefethen, J.M. and Dow, R.L., 1960. Some features of modern beach sediments. J. Sediment. Petrol., 30, 589-602.
- Trowbridge, A.C., 1911. The terrestrial deposits of the Owens Valley, California. J. Geol., 19, 709-747.
- Twelvetrees, W.H., 1908. Tyenna to Gell River. Tasm. Parl. Pap., 13, 25-33.
- _____, 1909. Western exploration: a report on a journey to the Gordon River. Tasm. Parl. Pap (1909), 25-35.
- Unrug, R., 1963. Istebna Beds - a fluxoturbidite formation in the Carpathian flysch. Ann. Soc. Geol. Pologne, 33, 49-92.
- Van Andel, T.H., 1958. Origin and classification of Cretaceous, Paleocene and Eocene sandstones of western Venezuela. Am. Assoc. Petrol. Geol. Bull., 42, 734-763.
- Van der Linden, W.J.M., 1963. Sedimentary structures and facies interpretation of some molasse deposits. Geologica Ultraiectina, No.12, 218pp.
- Van der Lingen, G.J., 1969. The turbidite problem. N.Z. J. Geol. Geoph., 12, 7-50.
- Van Straaten, L.M.J.U., 1954a. Sedimentology of recent tidal flat deposits and the Psammites du Condroz (Devonian). Geol. en Mijnb., 16, 25-47.
- _____, 1954b. Composition and structure of Recent marine sediments in the Netherlands. Leidse. Geol. Mededel., 19, 1-110.
- _____, 1960. Some recent advances in the study of deltaic sedimentation. L'pool Manch'r Geol. J., 2, 411-442.
- _____, 1961. Sedimentation in tidal flat areas. Jour. Alberta Soc. Petrol. Geol., 9, 203-226.
- Walker, R.G., 1965. The origin and significance of the internal sedimentary structures of turbidites. Proc. Yorks. geol. Soc., 35, 1-32.

- Walker, R.G., 1966a. Shale Grit and Grindslow Shales: transition from turbidite to shallow water sediments in the Upper Carboniferous of northern England. J. Sediment. Petrol., 36, 90-114.
- _____, 1966b. Deep channels in turbidite-bearing formations. Am. Assoc. Petrol. Geol., Bull., 50, 1899-1917.
- _____, 1967a. Upper flow regime bed forms in turbidites of the Hatch Formation, Devonian of New York State. J. Sediment. Petrol., 37, 1052-1058.
- _____, 1967b. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. Ibid., 37, 25-43.
- _____ and Sutton, R.G., 1967. Quantitative analysis of turbidites in the Upper Devonian Sonyea Group, New York. Ibid., 37, 1012-1022.
- Walton, E.K., 1967. The sequence of internal structures in turbidites. Scott. J. Geol., 3, 306-317.
- Ward, L.K., 1909. Contribution to the geology of Tasmania - Precambrian rocks. Pap. Roy. Soc. Tasm., 129-156.
- Waterhouse, J.B. and Bradley, J., 1957. Redeposition and slumping in the Cretaceous-Tertiary strata of southeast Wellington. Trans. Roy. Soc. N.Z., 84, 519-548.
- Wentworth, C.M., 1967. Dish structure, a primary sedimentary structure in coarse turbidites (Abstract). Am. Assoc. Petrol. Geol. Bull., 51, 485.
- Whitten, E.H.T., 1966. Structural Geology of Folded Rocks. Rand McNally, Chicago. 633pp.
- Williams, E., 1967. An outline of the geology of Tasmania, Bull. Geol. Surv. Tasm., 50, 9-26.
- Wolman, M.G. and Leopold, L.B., 1957. River floodplains: some observations on their formation. U.S. Geol. Surv. Prof. Pap. 282C, 87-107.
- Wood, A. and Smith, A.J., 1959. The sedimentation and sedimentary history of the Aberystwyth grits (Upper Llandoveryan). Quart. J. Geol. Soc. Lond., 114, 163-195.

APPENDIX A

ORDOVICIAN STRATIGRAPHY

INTRODUCTION

The author's mapping in the Florentine Synclinorium has necessitated considerable revision and expansion of the Ordovician stratigraphic nomenclature, and a brief outline of the revised sequence is presented herein. References quoted have already been listed in the main bibliography of the thesis, and will not be repeated here. Map co-ordinates refer to the Lands and Surveys Wedge Sheet.

The terminology applied to the Tasmanian Ordovician sequences has been a matter of some confusion and debate for many years, and some anomalies still exist. Carey (1947) recognized that the various Lower Palaeozoic conglomerate-limestone sequences in western and northwestern Tasmania were essentially all the same age (i.e. Ordovician) and could be thought of in terms of several type formations. Hills and Carey (1949) proposed the term "Junee Group" for this generalized sequence, following the work of Lewis (1940) in the Tim Shea-Maydena (formerly Junee) area, and defined it as consisting of five formations, viz. Jukes Breccias and Conglomerates at the base, followed by West Coast Range Conglomerate, Caroline Creek Sandstones and Shales, Gordon River Limestone, and Crotty Sandstones at the top.

The Crotty Formation has since been shown to be Silurian, and is now included in the Eldon Group, but the remainder of the succession with only slight modifications is still quoted, and the "Junee Group" is still used in this sense by some workers. Thus Banks (1962b, p.147), in a review of the Ordovician, states: "The Ordovician rocks of Tasmania are known as the Junee Group, which may be defined as consisting of the following formations or their correlates: Fenestella Shale at the top, Gordon Limestone (Lower to Upper Ordovician), Florentine Valley Mudstone, Caroline Creek Sandstone, Owen Conglomerate, Jukes Breccia." It will be noted that only two of the type formations (Gordon Limestone and Florentine Valley Mudstone) are from the type area of the "Junee Group" (i.e. Florentine Synclinorium), the others being from west coast and

northwest areas. The lack of a single complete Ordovician section in which all the formations are named and defined has been a major factor in the nomenclature disagreements.

The Tasmanian Department of Mines has adopted a somewhat different policy with respect to Ordovician nomenclature in their regional mapping programme. Recognizing the variations in the formations, particularly the lower ones, and the fact that deposition of these units may not have been continuous either spatially or temporally from one area to another, there has been a tendency to use local formation and group names for each basin or area. Thus Williams (in Jennings et al., 1967) regards the term "Junee Group" as applying only to the Florentine Valley sequence. This trend must continue as mapping of the state proceeds, and there seems little point in retaining a "Junee Group" in which most of the type formations occur outside the type area.

The present author's work in the Florentine Synclinorium, which may be considered as including the type area of the original "Junee Series" of Lewis (1940), suggests that the Ordovician section in this area is more complete and better exposed than any other section in Tasmania, and should thus serve as a basis for correlation throughout the state. A new terminology is proposed which incorporates those elements of the old terminology which are applicable, and is outlined below.

The basal conglomerates are given formation status (Reeds Conglomerate), as are the laterally-equivalent marine and non-marine sandstones around the eastern side of the synclinorium (Tim Shea Sandstone). The overlying sequence of marine sandstones and siltstones is given group status (Florentine Group), since three formations have been mapped in the northern Rasselas area and also appear to be present in the southern areas. The stratigraphy of the overlying limestone sequence is modified from that originally proposed by Corbett (1964). The limestones are given group status (Gordon Group), to consist of six formations, of which the upper four formations are included in the Benjamin Sub-Group because of mapping difficulties due to lack of outcrop of the thinner siltstone units. The total sequence will be known as the Junee Super-Group, and the terminology is shown in Fig. A1.

TERMINOLOGY OF JUNEE SUPER - GROUP

JUNEE SUPER - GROUP	GORDON GROUP	BENJAMIN SUB - GROUP	WESTFIELD SILTSTONE
			DAWSON LIMESTONE
			LORDS SILTSTONE
			PARKER LIMESTONE
			CASHIONS CREEK LIMESTONE
	FLORENTINE GROUP		WHERRETS CHERT MEMBER
			KARMBERG LIMESTONE
			TIMBS SANDSTONE
			SQUIRREL CREEK SILTSTONE
			CHAPLIN SANDSTONE
			Upper sandstone member
			REEDS CONGLOMERATE
			TIM SHEA SANDSTONE

FIGURE A1

The various units are defined below in accordance with the Australian Code of Stratigraphic Nomenclature.

DEFINITIVE STRATIGRAPHY

JUNEE SUPER-GROUP

Synonymy: Junee Series - Lewis, 1940; Junee System - Carey, 1947; Junee Group - Hills and Carey, 1949, and Banks, 1962b. Older terms are listed by Banks (1962b) and Smith (1957).

Derivation: The old township of Junee, now known as Maydena.

Type area: Florentine Synclorium (this thesis), including Denison Range, Tim Shea and Florentine Valley sections (see Fig. 2 of thesis, also Appendix C map).

Thickness: Maximum of about 14,000 feet (4200 m) in Denison Range-Rasselas Valley area.

Age and relationships: Upper Cambrian to Upper Ordovician; mostly conformable on Upper Cambrian sequences (Denison Group, Adamsfield Beds), otherwise unconformable on older rocks; apparently conformable and transitional with overlying Siluro-Devonian Eldon Group.

Elements: Consists of Reeds Conglomerate and Tim Shea Sandstone at base, followed by Florentine Group followed by Gordon Group.

REEDS CONGLOMERATE

Derivation: Reeds Peak (762N, 425E), highest point on Denison Range.

Type section: Crest and eastern slopes of Denison Range; well exposed in many other areas.

Lithology: Pebble- to boulder-grade siliceous orthoconglomerates with interbedded conglomeratic sandstones; usually red to purplish in colour. Further details given in thesis.

Thickness: Varies greatly in thickness (see Fig. 6 of thesis), from maximum of about 5200 feet (1560 m) near Reeds Peak to a few hundred feet or less in some areas.

Age: Virtually unfossiliferous, lies between fossiliferous Upper Cambrian and Lower Ordovician beds; could be largely Upper Cambrian.

Relationships: Probably interfingers to west with Tim Shea Sandstone (partly marine) and appears to be laterally equivalent to that formation; probably also interfingers basally with Denison Group and Adamsfield Beds, and at the top with marine sandstones of Florentine Group.

Palaeogeographic interpretation: Mainly alluvial fanglomerates deposited along margin of Tyennan Geanticline in down-faulted basins; has thin intercalations of shallow marine sandstone in a few places (Thumbs, Clear Hill) indicating proximity to shoreline.

Elements: Has an upper sandstone member in some areas, particularly on Stepped Hills, Denison Range, Battlement Hills, and Saw Back Range, but this unit is not given official status at this stage because of its apparent lateral variability.

TIM SHEA SANDSTONE

Synonymy: Tim Shea conglomerate and grits - Twelvetrees, 1908; Tim Shea Conglomerate - Opik, 1951, and Banks, 1962b.

Derivation: Tim Shea, a cuesta at the southern end of the Florentine Valley.

Type section: Lateral variability makes any single section inadequate - typical sections exposed at Tim Shea and on the cuestas to the southwest of this.

Lithology: At Tim Shea consists mainly of cross-bedded red sandstones, conglomeratic sandstones, some fine conglomerates, with a unit of red mudstone towards the base and a basal dolomitic breccia. Further west, near Mt. Mueller, the formation consists of zones of red, cross-bedded sandstone up to 100 feet or so thick alternating with zones of grey, flat-bedded sandstone with abundant worm burrows and very rare gastropods. Further details are given in main thesis.

Thickness: Varies greatly, from a maximum of almost 1000 feet (300 m) at Tim Shea and near Mt. Mueller, to a few hundred feet or less in some areas.

Age: Upper Cambrian - Lower Ordovician. The presence of Staurograptus diffusus (age Lancefieldian) in beds of the Florentine Group several hundred feet above the top of this formation suggests the Tim Shea Sandstone could be largely Upper Cambrian.

Relationships: Laterally equivalent to Reeds Conglomerate, with which it probably interfingers; probably partly equivalent to Adamsfield Beds also.

Palaeogeographic interpretation: Interfingering marine (grey) and non-marine (red) sandstones of coastal plain area seaward of large alluvial fans. Large variations in thickness suggest some tectonic control of deposition, e.g. the thickening near Mt. Mueller appears to correspond to a small Cambrian trough in this area.

FLORENTINE GROUP

Synonymy: Florentine Valley Mudstones - Etheridge, 1904 (in part); Junee mudstone - Lewis, 1940 (in part); Florentine Valley Mudstone - Banks, 1962b (in part). Encompasses Florentine Valley Mudstone plus "Caroline Creek Sandstone" of Banks, 1962b.

Type area: Florentine Synclinorium (this thesis); alternative sections available at Squirrel Creek (in the northern Rasselas Valley), also on Gordon Road near The Needles, and on Florentine Valley Road at The Gap. Formations have so far been defined only in the northern Rasselas area, and further work is required to link this section with those in the southern areas.

Age: Lower Ordovician (probably encompasses most of this epoch).

Palaeontology: The group contains an abundant marine fauna of trilobites, brachiopods, and gastropods in most areas, and graptolites, ostracods and other forms also occur. Worm burrows are commonly abundant in the sandstones. A number of fossil collections have been described from these rocks ("Florentine Valley Mudstone") from the Tim Shea-Maydena area (Etheridge, 1904;

Kobayashi, 1936, 1940; Brown, 1948; Opik, 1951; Singleton in Banks, 1962b; Thomas, 1960; Banks, 1962b; Corbett, 1964) but these have not yet been related to the formations proposed herein.

Relationships: Conformably overlies Reeds Conglomerate and Tim Shea Sandstone; conformably overlain by Gordon Group.

Formations: Chaplin Sandstone at base, Squirrel Creek Siltstone, Timbs Sandstone at top. Three similar formations can be mapped around the Tim Shea area (see Fig. A2), but these have not been named pending lithological and palaeontological correlation with the northern Rasselas section.

CHAPLIN SANDSTONE

Derivation: Chaplin Creek, a tributary of the Gordon River draining the northern end of the Denison Range (see Fig. 5 of thesis).

Type section: Lowermost eastern slopes of Denison Range in vicinity of Chaplin Creek and further north.

Lithology: White to grey quartzose sandstone with minor green siltstone; mostly flat bedding; worm tubes and gastropods abundant in most areas; glauconitic bands present in places.

Thickness: About 500 feet (150 m).

Palaeontology and age: Contains gastropods in nearly all areas, and rare trilobites in places, apart from the characteristic abundant worm burrows. Fossil collections have not yet been made from the type area, although Mr. M.R. Banks has tentatively identified a gastropod similar to Raphistoma sp. from this area. A collection of gastropods from a correlate of this formation in a track cutting on the southern tip of the Ragged Range includes forms similar to Ophileta sp. and Raphistoma sp., and Mr. Banks suggests a Lower Ordovician age. Probable correlates of the formation along the Gordon Road contain Lecanospira sp., again indicating a Lower Ordovician age. Further palaeontological work is obviously required.

Relationships: Conformably overlies the upper sandstone member of the Reeds Conglomerate, and probably interfingers with this unit

since the member contains horizons of similar grey, flat-bedded sandstone in places; conformably overlain by Squirrel Creek Siltstone.

Remarks: A similar sandstone unit overlies the Tim Shea Sandstone in most areas.

SQUIRREL CREEK SILTSTONE

Derivation: Squirrel Creek, a tributary of the Gordon River at the northern end of the Denison Range (see Fig. 5 of thesis).

Type section: Bed of Squirrel Creek just east of the Denison Gap, between 4271E and 4275E.

Lithology: Interbedded grey to yellow calcareous siltstones, fine sandstones, and impure modular limestones. Weathered outcrops have characteristic pock-marked appearance due to decomposition of the calcareous nodules.

Thickness: About 500 feet (150 m).

Palaeontology and age: Contains an abundant fauna of trilobites, brachiopods and gastropods similar to the collections which have been described from the Tim Shea area. Detailed palaeontological work on the formation in the type area has not yet been done - a small collection from here has been made by the author. The fossils include forms identified from the Tim Shea area, and indicate a Lower Ordovician age. The Staurograptus diffisus specimen identified by Dr. P.G. Quilty from near The Needles is from calcareous rocks which are possibly correlatable with this formation, suggesting a Lancefieldian age.

Relationships: Conformable with underlying and overlying formations.

Remarks: The formation characteristically forms a topographic low area between the sandstone ridges on either side, and is thus easily mappable once its presence is established in creek or road sections. A similar unit of siltstone and impure limestone forms a low area around the southeastern part of the synclinorium, but correlation is difficult in the area between the Saw Back Range

and the Denison Range. Excellent exposures of this siltstone-limestone unit (Fig. A2) occur on the Gordon Road just west of Tim Shea, and on the Florentine Valley Road at The Gap.

TIMBS SANDSTONE

Derivation: Timbs Ridge (7670N, 4275E), a low ridge near the foot of the northeastern end of the Denison Range (see Fig. 5 of thesis).

Type section: Squirrel Creek where it cuts through Timbs Ridge.

Lithology: Grey, green and buff-coloured quartzose sandstones with lesser siltstones; cross-bedded in places; glauconitic horizons common.

Thickness: About 1000 feet (300 m).

Palaeontology and age: Contains a sparse marine fauna of brachiopods, trilobites and gastropods, but no systematic collecting has yet been done. Relationships to underlying and overlying formations suggest it is Lower Ordovician, since base of overlying Gordon Group is Upper Canadian.

Remarks: The formation forms a distinctive low ridge around the Battlement Hills and the northern Denison Range (see Appendix C map). Further south, between The Thumbs and Tim Shea, the upper part of the Florentine Group consists of a thick sequence of interbedded sandstones, cherts, siltstones and minor impure limestones which is probably equivalent to this formation. The chert forms a residual white gravel on ridges in this area (Fig. A2).

GORDON GROUP

Synonymy: Gordon Limestone - Banks, 1962b. The previous nomenclature of this formation has been dealt with by Banks (1962b), and Smith (1957).

Type area: Florentine Valley (Fig. A2).

Thickness: Ranges from about 7000 feet (2100 m) in the Vale of Rasselas near the Denison Range, to about 3500 feet (1050 m) near The Thumbs. In the Florentine Valley the thickness is of the order of 4000-6000 feet (1200-1800 m).

SOLID GEOLOGY OF THE FLORENTINE VALLEY

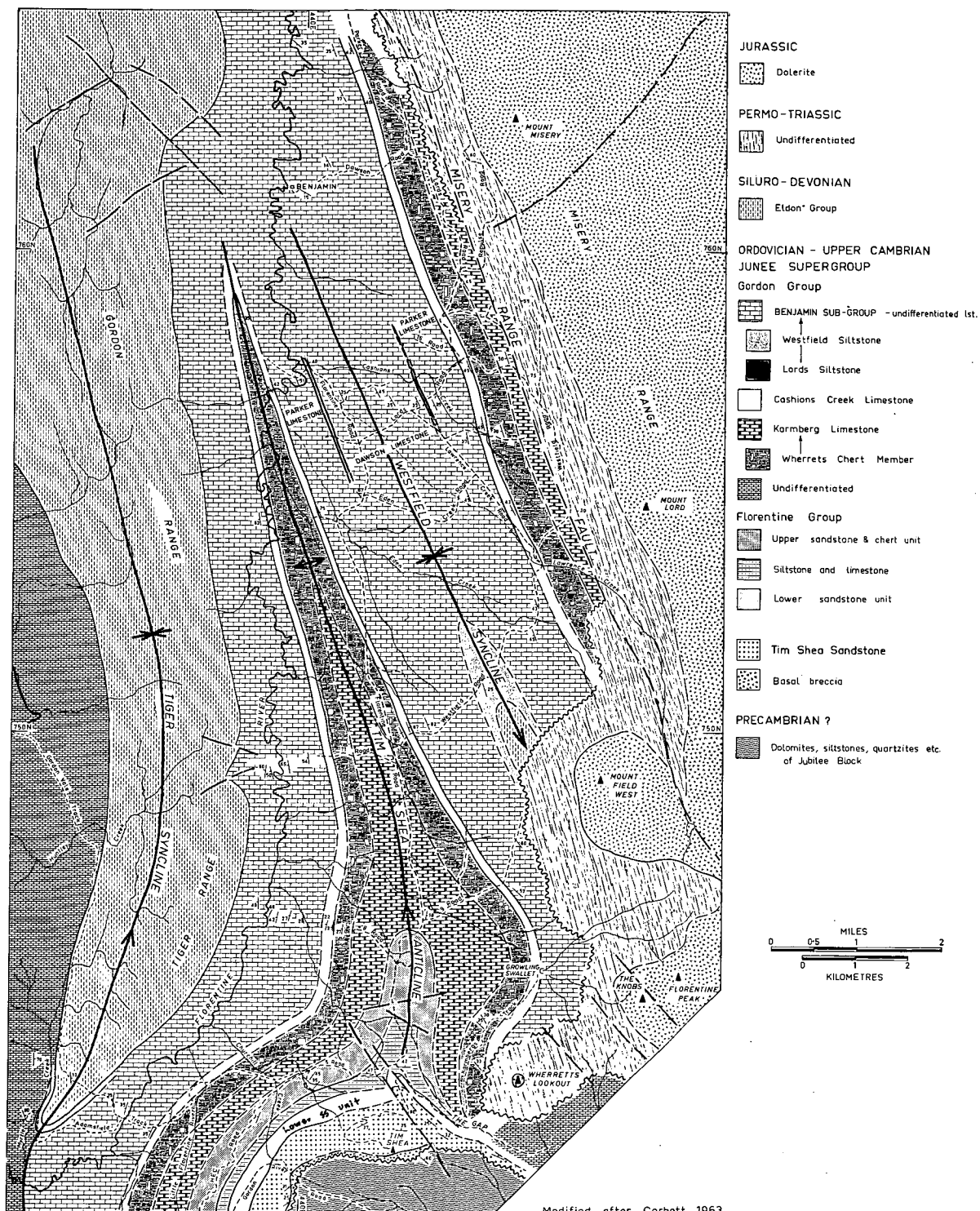


FIGURE A2

Lithology: Consists largely of dark grey, fine-grained marine limestone, but includes other types such as biocalcarenites, dolomitic limestone, cherty limestone, algal limestone, impure nodular limestone, and brown limonite-rich limestone. A thin unit of terrigenous siltstone and fine sandstone (Lords Siltstone of this thesis) occurs in the middle part of the section in the Florentine Valley, and the top of the sequence is formed by a similar unit (Westfield Siltstone of this thesis). A thick unit of coarse quartzose sandstone occurs in the middle part of the sequence in the Rasselas Valley (well exposed in Squirrel Creek), and is probably equivalent to the Lords Siltstone. North of Battlement Hills much of the lower half of the sequence consists of coarse quartzose grits and fine conglomerates which form a prominent ridge (see Appendix C map). These grits contain worm burrows and other poorly preserved fossils but have not yet been dated. No detailed petrological work on any of the limestone sequence has yet been attempted.

Palaeontology: Contains a typical Ordovician shelly fauna, including corals, brachiopods, trilobites, bryozoans, gastropods, cephalopods, stromatoporoids, algae, etc. (Banks, 1961, 1962b), with affinities to East Asia, North America, the Baltic, and New South Wales. A composite section showing some of the typical forms found in the Florentine Valley sequence is given in Fig. A3. No comprehensive study of the palaeontology of the group has yet been attempted.

Age: A cephalopod fauna from the base of the group at Adamsfield is probably Upper Canadian (Teichert and Glenister, 1953), while the fauna from the Westfield Siltstone at the top of the group is probably Upper Ordovician (M.R. Banks, pers. comm.). A poorly-preserved brachiopod fauna from sandstones overlying the Westfield Siltstone at the Tiger Range and near Mt. Field West also appears to be Upper Ordovician (M.R. Banks, pers. comm.). Thus the group ranges from Lower Ordovician to Upper Ordovician.

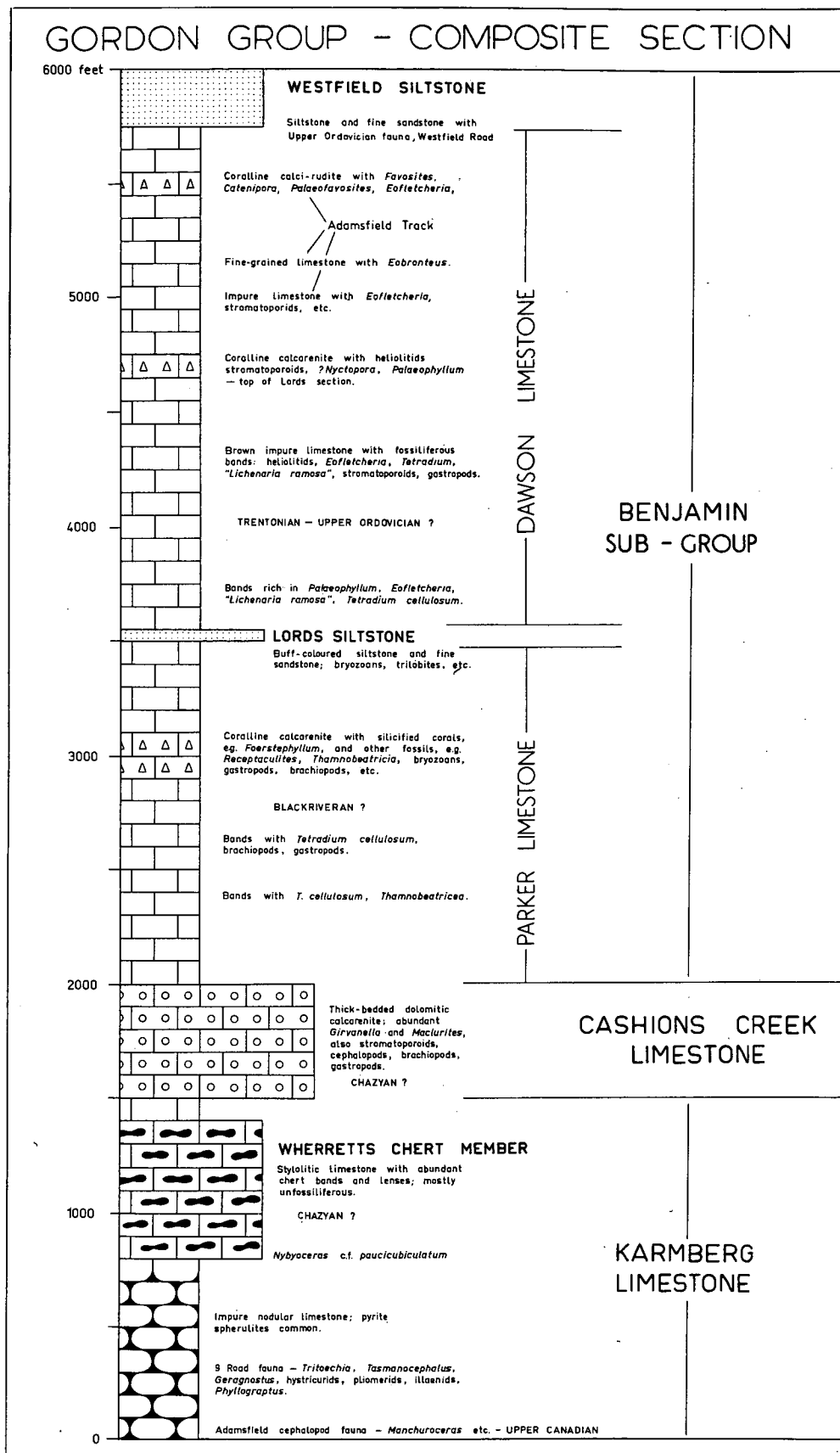


FIGURE A3

Relationships: Conformably overlies Florentine Group and is conformably overlain by Eldon Group.

Palaeogeographic interpretation: Predominance of dark micritic limestone and paucity of clean bioclastic types possibly suggests a fairly restricted (barred-basin ?) marine environment rather than open shelf conditions (Dr. J. Weiser, pers. comm.). To the west, in the Rasselas Valley, there is a much higher proportion of terrigenous material (much of the limestone here is impure and nodular) in the lower half of the sequence, suggesting proximity to a shoreline to the west (Tyennan Geanticline ?). The large thickness of quartzose sandstone in the Battlement Hills area indicates that near-shore conditions prevailed here for a considerable period, possibly into the Middle Ordovician.

Elements: Subdivided into six formations in the Florentine Valley. The upper four formations are included in the Benjamin Sub-Group, since the two thin siltstone formations are only mappable in areas of good outcrop. The subdivision is shown in Figs. A1, A3.

Remarks: Because of their high solubility the Ordovician limestones characteristically form broad, flat-floored, poorly-drained solution valleys usually covered with extensive superficial gravels. The Vale of Rasselas is typical of the topographic expression of the group in western Tasmania. Because of the lack of outcrop under such conditions, stratigraphy is usually difficult or impossible. Exposure is much better than normal in the Florentine Valley because the old drainage pattern has been disrupted by river capture, and the present Florentine River and tributaries have dissected much of the limestone surface and removed much of the superficial cover. The stratigraphic subdivision of the limestone sequence outlined herein is based very largely on the author's original work in the Florentine Valley (Corbett, 1964), and the purpose of the present account is mainly to define the units. The only additional work has been the tracing of the formations into the southern Rasselas Valley (see Appendix C map), and some revision of the terminology and the boundary with

the Florentine Group. Further descriptive details are given in Corbett (1964).

KARMBERG LIMESTONE

Derivation: Karmberg's Track, an old track leading from The Gap around the eastern side of the valley (Fig. A2).

Type section: Best exposures are on lower northern slopes of Wherretts Lookout and along 9 Road and its branches to the east (Fig. A2).

Lithology: Lower part of formation consists mainly of impure nodular limestone, such as exposed at the 9 Road junction, usually richly fossiliferous. Large pyrite spherulites are present in many places within this part. Upper half consists mainly of chert-rich limestone (Wherretts Chert Member), but the abundance of chert decreases markedly to the west.

Thickness: Approximately 1500 feet (450 m).

Palaeontology and age: Lower part has abundant marine fauna, e.g. 9 Road junction fauna (Fig. A3). Correlate of this formation at base of sequence at Adamsfield has Upper Canadian cephalopod fauna (Teichert, 1947; Teichert and Gelnister, 1953). Upper part poorly fossiliferous, may be largely Chazyan.

Members: Wherretts Chert Member.

WHERRETTS CHERT MEMBER

Derivation: Wherretts Lookout, near Tim Shea (Fig. A2).

Type section: Lower northwestern slopes of Wherretts Lookout.

Lithology: Dark grey limestone containing up to 50% chert as irregular beds, lenses and nodules.

Thickness: About 600 feet (180 m).

Palaeontology and age: Poorly fossiliferous for most part, although silicified brachiopods occur in chert in places. Single specimen of Nybyoceras paucicubiculatum collected from near base of member at type area. This species also occurs at Railton, and is thought to be Chazyan (Teichert and Glenister, 1953).

Remarks: Forms low ridges covered with residual chert gravel on valley floor. Amount of chert apparently diminishes to west, and member is not distinguishable in southern Rasselas area.

CASHIONS CREEK LIMESTONE

Derivation: Cashions Creek (7577N, 4410E), a tributary of the Florentine River (Fig. A2).

Type section: Well exposed in many areas, particularly where Cashions Creek is crossed by an easterly branch road of Lawrence Creek Road (7563N, 4437E) and for several miles north of this.

Lithology: Mostly thick-bedded dolomitic limestone containing very abundant Girvanella colonies; crops out strongly in most areas because less soluble than overlying and underlying limestone.

Thickness: About 500 feet (150 m).

Palaeontology: Characterized by abundant Girvanella; the large gastropod Maclurites is a common associate. Other fossils include stromatoporoids, cephalopods (including Orthonybyoceras tasmaniense), brachiopods, trilobites and gastropods. According to Banks and Johnston (1957), three species of Girvanella are present, while the species of Maclurites most closely resembles M. magnus, the type Chazyan species of North America.

BENJAMIN SUB-GROUP

Derivation: The old settlement of Benjamin in central part of Florentine Valley (Fig. A2).

Lithology: Includes all the sequence between Cashions Creek Limestone below and the Eldon Group above; and thus forms a mappable unit when the siltstone formations are not exposed (Fig. A2).

Elements: Consists of four formations, viz. Parker Limestone at base, Lords Siltstone, Dawson Limestone, and Westfield Siltstone at top. These are defined below.

PARKER LIMESTONE

Derivation: Parker logging division, which extends from Dawson Road north (Fig. A2).

Type section: Best exposed in area of 16th Road, between Lords Road and Cashions Creek.

Lithology: Limestone of various types, mainly dark and fine-grained with thin fossiliferous bands. A distinctive horizon of coralline calcarenite occurs in the middle part of the formation (Fig. A3) and is exposed just west of the end of 16 Road and also near the junction of Eden Creek Road and Lawrence Creek Road.

Thickness: About 1500 feet (450 m).

Palaeontology and age: Some of the characteristic fossils are listed in Fig. A3. Bands of Tetradium cellulorum are common in the lower part. The calcarenite zone contains silicified colonies of Foerstephyllum sp. as well as many other types. The fossils suggest a Blackriveran age, but detailed palaeontology is obviously required.

Remarks: Formerly called "Lower Limestone Member" by Corbett (1964).

LORDS SILTSTONE

Derivation: Lords Road, central Florentine Valley (Fig. A2).

Type section: Best exposed on main Florentine Road about 200 yards east of the bridge over the Florentine River (Fig. A2).

Lithology: Buff-coloured micaceous siltstone and quartzose fine sandstone.

Thickness: About 50 feet (15 m).

Palaeontology and age: Fauna includes trilobites, brachiopods, abundant bryozoans, cystoids and ostracods. Some tentative identifications are given by Corbett (1964), and suggest a Middle Ordovician age, possibly Lower Trentonian.

Remarks: Formerly called "Florentine Bridge Siltstone Member" by Corbett (1964).

DAWSON LIMESTONE

Derivation: Dawson Road, an old road formerly providing access to Benjamin (Fig. A2).

Type section: No single complete section yet available. Most of the unit is exposed in the core of the Westfield Syncline in the triangle between Cashions Creek, Lords Road and Florentine Road. The upper part of the formation is poorly exposed along the Adamsfield Track between the Florentine River and the southern tip of the Tiger Range.

Lithology: Limestone of various types, mainly dark and fine-grained with thin fossiliferous bands. Includes two zones of coralline calcarenite, one in the Lords Road area and one at the Tiger Range near the top of the formation.

Thickness: 2000-2500 feet (600-750 m).

Palaeontology and age: Some of the typical fossils are listed in Fig. A3. Eofletcheria sp. is very characteristic. Banks (1957) reports Favosites sp., Palaeofavosites sp., Catenipora sp. and others from the calcarenite-calcirudite at the Tiger Range. The fossils suggest a Trentonian to Upper Ordovician age at this stage.

Remarks: Formerly called "Upper Limestone Member" by Corbett (1964).

WESTFIELD SILTSTONE

Derivation: Westfield Road, southern Florentine Valley (Fig. A2).

Type section: Best exposed on Westfield Road 1½ miles east of Florentine Road in core of Westfield Syncline, and along a forestry track which leads south from here. Also exposed on the Adamsfield Track 100 yards southeast of Myrtle Creek.

Lithology: Buff-coloured siltstones and fine sandstones, with some interbedded coarse sandstones.

Thickness: About 500 feet (150 m) but difficult to measure because base is seldom exposed.

Palaeontology and age: Siltstones contain fairly rich fauna of bryozoans, brachiopods, trilobites, pelecypods, and crinoid fragments. Preliminary work by Mr. M.R. Banks suggests an Upper Ordovician age.

Remarks: Referred to as "siltstones and sandstones overlying Gordon Limestone" By Corbett (1964).

APPENDIX B

CATALOGUE OF ROCK SPECIMENS AND THIN SECTIONS

Specimens are arranged in stratigraphic order as used in Chapter 1 of the thesis. Where thin sections have been cut, these have the same number as the hand specimen, and are indicated by a +. Outsize thin sections are indicated by a *. Map co-ordinates refer to the Lands and Surveys Wedge Sheet.

Specimen Number	Rock Type and Locality	Thin Section
<u>OLDER PRECAMBRIAN OF TYENNAN GEANTICLINE</u>		
37121	Crenulated pyritic quartz-schist beneath Trial Ridge Beds, bed of Kindling Creek at north end of Trial Ridge.	
<u>YOUNGER PRECAMBRIAN OF JUBILEE BLOCK</u>		
37122	Lithic greywacke, core of Needles Anticline.	+
37123	Laminated dolomite above greywacke, core of Needles Anticline.	
37124	Dolomite from "Humboldt Slate and Dolomite", small quarry on Gordon Road, east flank of The Needles.	
37125	Banded siltstone from "Humboldt Slate and Dolomite" Gordon Road, east flank of The Needles.	
37126	Silicified breccia within dolomite, old South Gordon Track, $\frac{1}{4}$ mile east of junction with Port Davey Track (see Appendix C map).	
37127	Banded chert, same locality as 37126.	
37128	White quartzite, anticlinal structure, South Gordon Track, 1 mile east of Port Davey Track.	
37129	Dolomite (conglomeratic), Port Davey Track, $\frac{1}{2}$ mile south of junction with South Gordon Track.	
<u>OLDER CAMBRIAN OF BOYD BASIN</u>		
37130	Sandstone, from conglomerates overlying Wedge River Beds, 3 miles west of Mt. Wedge on Gordon Road.	+
37131	Green chloritic sandstone, immediately overlying above conglomerates.	+
37132	Graded sandstone with load casts, from turbidite-siltstone sequence on Gordon Road just northwest of Mt. Wedge.	+
37133	Laminated siltstone, as for 37132.	+
37134	Thin sandy turbidite, as for 37132.	+

Specimen Number	Rock Type and Locality	Thin Section
37135	Perfect graded turbidite with convolute lamination (other half is numbered 35863), as for 37132. <u>Greywacke-chert-mudstone sequence of Boyd Basin</u>	
37136	Greywacke, near headwaters of Boyd River (west branch), south of Gordon Road.	+
37137	Greywacke, just west of serpentinite contact on Gordon Road, southern end Saw Back Range.	+
37138	As for 37137.	+
37139	Greywacke, Gordon Road, just north of Mt. Wedge (7289N, 4245E).	+
37140	As for 37139.	
37141	Fresh greywacke, just west of IXL Sawmill, Gordon Road near Mt. Wedge.	
37142	Banded chert, small quarry with "CHERT" sign, Gordon Road north of Mt. Wedge.	+
37143	Greywacke, just east of serpentinite contact, Gordon Road, southern Saw Back Range.	+
37144	Red mudstone (weathered), Gordon Road, just north of Mt. Wedge.	
37145	Limestone (age unknown) at serpentinite contact, Gordon Road.	
37146	As for 37145.	+
37147	Red chert, Adams River, just west of north end of Ragged Range.	
37148	Siltstone, as for 37147.	+
37149	Red mudstone, bombardier track, north end of Ragged Range.	
	<u>Igneous rocks of Boyd Basin</u>	
37150	Andesite (?) lava, ½ mile west of Boyd River crossing on Gordon Road (7287N, 4294E).	
37151	Andesite (?) lava, as for 37150.	+
37152	Fine-grained basic rock (microgabbro ?), Gordon Road near Mt. Wedge (7291N, 4267E).	+
37153	Gabbro, Gordon Road, west branch of Boyd River (7279N, 4287E).	
37154	As for 37153.	+
	<u>CAMBRIAN ROCKS FROM NEAR MT. MUELLER</u>	
37156	Tuff, boulder in creek bed (7336N, 4368E).	+
37157	Fine-grained basic igneous rock, creek bed as for 37156.	
37158	Dolerite (?), cliffs about 200 yards upstream of 37156.	+
37159	Altered siltstone at contact with above dolerite.	+
37160	Greywacke, adjacent to siltstone of 37159.	+

Specimen Number	Rock Type and Locality	Thin Section
37161	Gabbro (?), about 100 yards upstream of 37159, on east bank.	+
37162	Feldspathic sandstone or tuff, as for 37161.	+
37163	Trachy-andesite (?), just upstream of 37161.	+
37164	Crystal tuff, as for 37163.	+
37165	Crystal tuff, as for 37163.	+
37166	Sandstone, just upstream of 37165.	+
37167	Fine greywacke, lower part of this creek near contact with Ordovician.	
37168	As for 37167.	+
<u>CAMBRIAN ROCKS FROM CLEAR HILL AREA</u>		
37170	Greywacke-conglomerate, south bank of Gordon River (7494N, 4236E).	+
37171	Black slate, as for 37170.	
37172	Greywacke sandstone, as for 37170.	+
<u>TRIAL RIDGE BEDS</u>		
37174	Sandy matrix of coarse conglomerate, upper marine part of Trial Ridge Beds (7609N, 4231E).	+
37175	Graded sandy turbidite, as for 37174. Thin sections from basal and upper parts.	++
37176	Sandy turbidite, oxidized to red colour, near unconformity beneath Singing Creek Formation, north end of Trial Ridge.	+
<u>ADAMSFIELD ULTRAMAFIC BODY</u>		
37178	Pyroxenite, from ridge NNE of Adamsfield (7421N, 4307E).	
37179	Chromite vein, cross-cuts banding in above ridge, as for 37178.	
37180	Quartzite from tectonic inclusion in serpentinite, NE corner of body (7423N, 4315E).	+
37181	Contact jasperoid rock, adjacent to quartzite of 37180.	+
37182	As for 37181.	+
37183	As for 37181.	+
37184	Altered serpentinitic breccia containing riebeckite amphibole, $\frac{1}{4}$ mile north 37180.	+
37185	As for 37184.	+
37186	As for 37184.	+
37187	Blue contact rock with riebeckite amphibole, small trench $\frac{1}{4}$ mile south of 37180.	+
37188	Serpentinite from Boyes Valley ultramafic body, Boyes River.	

Specimen Number	Rock Type and Locality	Thin Section
<u>MIDDLE OR UPPER CAMBRIAN SEQUENCE ON SCOTTS PEAK ROAD</u>		
37190	Red dolomitic siltstone, 1½ miles south of Gordon Road junction, near small road metal quarry (7271N, 4339E).	+
37191	Fine conglomerate, as for 37190.	+
37192	Serpentinitic conglomerate, road metal quarry as for 37190.	+
37193	Talc pebble with chromite grains, in conglomerate as for 37192.	+
37194	Greywacke turbidite, quarry as for 37190.	+
37195	Dolomitic siltstone, as for 37190.	+
37196	Serpentinitic conglomerate, as for 37192.	
37197	As for 37196, with large fragment of serpentinite.	
37198	Greywacke turbidite, quarry as for 37192.	+
37199	Dolomitic sandstone, 200 yards west of above quarry.	
37200	Dolomitic conglomerate, as for 37199.	
37201	Dolomitic breccia, in large fault block, Scotts Peak Road, about ½ mile south of above quarry.	
37202	Coarse greywacke, overlies dolomite of 37201.	+
37203	Diorite, adjacent to serpentinite outcrops at "Serpentine Creek", Scotts Peak Road.	+
<u>Boulders of Older Cambrian igneous rocks from conglomerates, Scotts Peak Road</u>		
37204	Dolerite (?) pebble in conglomerate, about 150 yards east of quarry of 37190.	+
37205	Basic igneous pebble, as for 37204.	
37206	Basic igneous pebble, as for 37204.	
37207	Pyroxenite boulder in creek gravel derived from conglomerates on ridge just west of Scotts Peak Road.	+
37208	Granophyre boulder, as for 37207.	+
37209	Granophyre (?) boulder, as for 37207.	+
37210	Quartz-diorite boulder, as for 37207.	+
37211	Basic to intermediate lava (?) pebble, as for 37207.	+
<u>ADAMSFIELD BEDS (Upper Cambrian)</u>		
37213	Serpentinitic sandstone near contact with serpentinite, old track along contact at head of Main Creek, about 200 yards north of open cut (7407N, 4317E).	
37214	Impure nodular limestone, 20 yards east of head of open cut	+
37215	Red serpentinitic conglomerate, interbedded with fine sandstones and shales, small trench about 100 yards east of open cut.	+
37215	Red sandstone from conglomerate bed of 37215 above	+

Specimen Number	Rock Type and Locality	Thin Section
37216	Sandstone with gastropods, upper part of Adamsfield Beds, track leading to reservoirs near top of ridge north of Adamsfield Track.	+
37217	Sandstone with chromite grains, southern reservoir (7413N, 4320E).	
37218	As for 37217.	+
37219	Coarse conglomerate with abundant chromite in matrix, also red jasperoid fragments, just east of reservoir of 37217.	
37220	Sandstone with parting lineation, northern reservoir (100 yards north of 37217).	+
37221	Coarse grey sandstone, northern reservoir.	+
37222	Red sandstone, top of northern reservoir section	+
<u>Correlates of Adamsfield Beds on Ragged Range</u> <u>(below Reeds Conglomerate)</u>		
37223	Basal cherty conglomerate, near bombadiers track, north end of Ragged Range	+
37224	Sandstone above basal conglomerate, 1 mile south of 37223.	+
37225	Limonitic fine sandstone, middle part of sequence towards south end of Ragged Range.	+
<u>SINGING CREEK FORMATION</u> (west flank of Denison Range)		
37227	Grey siltstone (Group A), near north end of Denison Range	+
37228	Group A siltstone, knoll below Staircase Rocks - shown as Plate 5B of thesis	+
37229	Siltstone with graded laminae, as for 37227 - shown as Plate 5A of thesis.	+
37230	Fossiliferous siltstone with mottled texture, lower western slope of Great Dome.	+
37231	Laminated fine sandstone (Group A), as for 37230.	+
37232	Coarse siltstone, as for 37230.	+
37233	Laminated siltstone, as for 37230.	+
37234	Fine sandstone, as for 37230.	
37235	Laminated siltstone, south flank of Flagstone Knoll	+
37236	Rippled sandy lamina in siltstone, west slope of Great Dome - see Plate 6A of thesis	+
37237	Papery dark siltstone, bed of Kindling Creek just northwest of Flagstone Knoll.	+
37238	Fossiliferous siltstone, spur above Boyes River near Flagstone Knoll.	
37239	Laminated siltstone from Slump No.14, west flank Flagstone Knoll (Plate 27 of thesis).	
37240	Calcareous fine sandstone from Slump No.3, west flank of Flagstone Knoll	

Specimen Number	Rock Type and Locality	Thin Section
37241	Very calcareous fine sandstone, middle reaches of Singing Creek.	+
35848	Thin fine sandstone with longitudinal ridges, cliffs near head of Singing Creek.	
35849	As for 35848.	
35850	Thin fine sandstone with longitudinal ridges on sole, south flank of Flagstone Knoll.	
<u>Group B. Thin-bedded sandstones</u>		
37243	Thin graded bed, south flank of Flagstone Knoll.	*
37244	Thin graded bed with rippled top, south flank Flagstone Knoll.	*
37245	Thin graded bed within conglomeratic zone (Fig. 10 of thesis), south flank Flagstone Knoll - bed immediately beneath Bed 1 of Fig. 10.	*
37246	Thin sandstone bed underlying that of 37245.	+
37247	Thin bed beneath 37246.	+
37248	Thin bed beneath 37247.	+
37249	Sandy lamina above Slump No.26, west flank Flagstone Knoll - shown in Plate 11B and Fig. 8 of thesis as bed A.	*
37250	Conglomeratic part of Bed B1 in Fig. 8 of thesis - overlies 37249. Modal analysis and grainsize distribution given in Table 1 and Fig. 9 of thesis.	+
37251	Bed B2 in Fig. 8 of thesis, overlies 37250.	*
37252	Thin sandstone part of Bed C near north end of Fig. 8 of thesis, overlies 37251.	*
37253	Conglomeratic bed overlying 37252.	
37254	Sandstone ball in slump sheet, west flank Flagstone Knoll.	+
37255	Bed with flute marks, rippled top, overlies Slump No.8, west flank Flagstone Knoll. Plate 14B of thesis.	+
37256	Sandstone from composite sequence of Plate 7B, west flank Flagstone Knoll. No. B3 in Table 1 and Fig. 9 of thesis.	+
37257	Lower half of bed with flute marks overlying Slump No. 25, west flank Flagstone Knoll. Used for modal and grainsize analysis - B2(a) and B2(b) are from base and top, in Table 1 and Fig. 9.	++
37258	Upper half of bed of 37257. B2(c) of Table 1 and Fig. 9 of thesis.	+
37259	Bed with load casts and rippled top, from cliffs near head of Singing Creek. B4 of Table 1 and Fig. 9 of thesis. Photomicrograph shown as Plate 18B.	++
37260	Sandstone ball in slump sheet, knoll below Staircase Rocks.	+

Specimen Number	Rock Type and Locality	Thin Section
37261	Sandstone bed, knoll below Staircase Rocks.	+
37262	As for 37261.	+
35858	Bed with longitudinal ridges and furrows, cliffs near head of Singing Creek. Plate 14A of thesis.	
35851	Upper part of 35858.	
35853	Sole of thin sandstone showing truncated ripple structure, west flank Flagstone Knoll.	
35857	Part of same specimen as 35853.	
35855	Sole of thin bed showing current crescents around worm tubes, south flank Flagstone Knoll.	+
35864	Sole of bed showing large oriented load casts, south flank Flagstone Knoll. Plate 15B of thesis.	
<u>Group C. Thick-bedded sandstones</u>		
37264	Sandstone from bed showing dish structure, crest of Flagstone Knoll. C1 in Table 1 and Fig. 9 of thesis.	+
37265	Sandstone with elutriation columns, crest of Flagstone Knoll. C2 in Table 1 and Fig. 9.	+
37266	Sandstone with parting lineation, upper part of bed, crest Flagstone Knoll. Thin sections cut perpendicular (No.1) and parallel (No.2) to parting lineation.	++
37267	Thinnest part of the central bed shown in Plate 21A of thesis - outcrop about 200 feet above head of Singing Creek on east side.	
37268	Basal part of bed with dish structure, shown in Plate 20 of thesis. Locality as for 37267.	+
37269	Middle part of same bed as 37268, near top of dish structure division.	+
37270	Upper part of same bed as 37268, in division of large-scale cross-bedding.	+
<u>Group D. Conglomerates</u>		
37272	Fine conglomerate, knoll below Staircase Rocks.	+
37273	Conglomeratic sandstone, crest of ridge above Boyes River (7609N, 4234E).	+
37274	Fine conglomerate, as for 37273. D3 of Table 1 and Fig. 9 of thesis.	+
37275	Lower part of conglomerate bed shown as Bed 1 in Fig. 10 and Plate 25. D1 of Table 1 and Fig. 9. South flank of Flagstone Knoll.	+
37276	Base of Bed 3 in Fig. 10, also shown in Plate 26A. D2 of Table 1 and Fig. 9.	+
37277	Sandy upper part of Bed 2 in Fig. 10. D4 in Table 1 and Fig. 9.	+
37278	Upper part of Bed 1 in Fig. 10, showing contact between coarse and fine strata.	+

Specimen Number	Rock Type and Locality	Thin Section
37279	Part of conglomerate bed with abundant intra-formational fragments of sandstone, west flank of Flagstone Knoll.	
	<u>Group E. Slump sheets</u>	
37281	Matrix of pebbly mudstone slump, south flank Flagstone Knoll.	+
37282	As for 37281.	+
37283	Sandy matrix of incoherent slump sheet, west flank Flagstone Knoll.	+
37284	Sandy matrix of Slump No.22, west flank Flagstone Knoll.	+
37285	Matrix of Slump No.20, as for 37284.	+
37286	Sandy matrix of Slump No.4, west flank Flagstone Knoll.	+
<u>GREAT DOME SANDSTONE (west flank of Denison Range)</u>		
37288	Lower part of bed showing bioturbation and pseudo-graded bedding, Transition Beds, Staircase Rocks.	++
37289	Upper part of same bed.	+
37290	Small pseudonodule of fine sandstone in silty sandstone, Transition Beds, Staircase Rocks.	*
37291	Siltstone from thick silt bed, Transition Beds, Staircase Rocks.	+
37292	Coarse sandstone from thick-bedded unit of Lithofacies A, near Bonds Craig.	+
37293	Lithofacies A sandstone, Staircase Rocks.	+
37294	Small pseudonodule of fine sand in coarse silt, probably Lithofacies B, Staircase Rocks.	
37295	Lithofacies B bioturbated sandstone with limonite veins, Staircase Rocks.	
37296	Lithofacies B bioturbated sandstone with U-shaped worm burrows, Staircase Rocks.	+
37297	Lithofacies C siltstone with pseudonodule of fine sandstone, Staircase Rocks.	*
37298	Lithofacies C siltstone from Cycle 1, Staircase Rocks. Plate 35B of thesis.	+
37299	Lithofacies C siltstone from Cycle 1, near base of siltstone unit.	
37300	Lithofacies C siltstone, as for 37298.	+
37301	Lithofacies D sandstone, from large "raft" of Cycle 1, shown in Plate 35B.	+
37302	Lithofacies E sandstone, containing gastropods, from Cycle 1, Staircase Rocks.	
37303	Lithofacies E sandstone with flat lamination, current crescents around worm tubes. Staircase Rocks.	

Specimen Number	Rock Type and Locality	Thin Section
37304	Lithofacies E sandstone with parting lineation, Staircase Rocks. Thin sections cut perpendicular (No.1) and parallel (No.2) to lineation.	++
37305	Sandstone with unusual polygonal ripple mark (interference ripples ?) from base of small channel structure in thin-bedded zone, upper part of Staircase Rocks section.	
37306	Glaucinitic sandstone with brachiopods, middle part of Great Dome Sandstone at Stepped Hills.	+
35856	As for 37303 - current crescents around worm tubes. Plate 37B of thesis.	

REEDS CONGLOMERATE

37308	Red sandstone near base of formation at Bonds Craig.	+
37309	Conglomeratic sandstone, Upper Sandstone Member, east of Bonds Craig.	+
37310	As for 37309.	+
37311	Grey sandstone from grey horizon in Upper Sandstone Member, northern end of Mt. Wright.	+
37312	Red sandstone unit in Upper Sandstone Member, 1 mile southwest of Mt. Wright.	+
37313	Red sandstone above Adamsfield Beds, ridge between Thumbs and Saw Back Range.	+
37314	Matrix of conglomerate, Gordon Road at southern end Saw Back Range.	+

TIM SHEA SANDSTONE

37316	Grey sandstone horizon, ridge just west of Mt. Mueller.	+
37317	Grey sandstone with worm tubes, as for 37316.	+
37318	Fine grey conglomerate, as for 37316.	+
37319	Fossiliferous sandstone near base of formation, overlying Cambrian rocks near Mt. Mueller.	
37320	Red conglomeratic sandstone, cutting on Gordon Road just west of Tim Shea.	
37321	Red sandstone from red horizon, ridge just north of west end of Mt. Mueller.	

FLORENTINE GROUP (Lower Ordovician)

37323	Grey sandstone with animal tracks, Chaplin Sandstone formation, north end of Denison Range.	
37324	Sandstone with gastropods, as for 37323.	
37325	Sandstone with gastropods, as for 37323.	

Specimen Number	Rock Type and Locality	Thin Section
37326	Chaplin Sandstone with animal tracks, as for 37323.	+
37327	Squirrel Creek Siltstone - nodular impure limestone, bed of Squirrel Creek, north end of Denison Range.	+
37328	Squirrel Creek Siltstone, calcareous siltstone, as for 37327.	+
37329	Sandstone with compound animal burrows, base of Timbs Sandstone in Squirrel Creek.	
37330	Timbs Sandstone - glauconitic sandstone, ridge north of Battlement Hills.	+
37331	Timbs Sandstone - glauconitic sandstone, bed of Squirrel Creek near Timbs Ridge.	+
<u>GORDON GROUP</u>		
37333	Coarse sandstone with fossils, ridge in Rasselas Valley just north of Battlement Hills.	
37334	Fine quartzose conglomerate, as for 37333.	+
37335	Fine conglomerate, as for 37333.	
37336	Quartzose sandstone, as for 37333.	+